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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 71, PP. 645-768, 35 FIGS. JUNE 1960

REVIEW OF PALEOMAGNETISM

BY ALLAN COX AND RICHARD R. DOELL

ABSTRACT

This review is an attempt to bring together and discuss relevant information concerning the magnetization of rocks, especially that having paleomagnetic significance. All paleomagnetic measurements available to the authors are here compiled and evaluated, with a key to the summary table and illustrations in English and Russian. The principles upon which the evaluation of paleomagnetic measurements is based are summarized, with special emphasis on statistical methods and on the evidence and tests for magnetic stability and paleomagnetic applicability.

Evaluation of the data summarized leads to the following general conclusions:

(1) The earth's average magnetic field, throughout Oligocene to Recent time, has very closely approximated that due to a dipole at the center of the earth oriented parallel to the present axis of rotation.

(2) Paleomagnetic results for the Mesozoic and early Tertiary might be explained more plausibly by a relatively rapidly changing magnetic field, with or without wandering of the rotational pole, than by large-scale continental drift.

(3) The Carboniferous and especially the Permian magnetic fields were relatively very "steady" and were vastly different from the present configuration of the field.

(4) The Precambrian magnetic field was different from the present field configuration and, considering the time spanned, was remarkably consistent for all continents.

RÉSUMÉ

Ce compte-rendu s'efforce de rassembler et de discuter les données utiles sur la magnétisation des roches, et en particulier celles qui ont un intérêt paléomagnétique. Toutes les mesures paléomagnétiques connues des auteurs sont rassemblées et évaluées (avec un indexe du tableau-résumé et des illustrations en anglais et en russe). Les principes sur les quels se base l'évaluation des mesures paléomagnétiques sont resumés, en appuyant sur les méthodes statistiques et sur les données et les tests qui révèlent la stabilité magnétique et l'utilisation possible en paléomagnétisme.

L'étude des données mène aux conclusions générales qui suivent:

1. Le champ magnétique moyen de la terre, de l'Oligocène à la période actuelle suit de très près ce qui résulterait de la présence d'un dipôle au centre de la terre, orienté parallèlement à l'axe de rotation actuel.

2. Les données paléomagnétiques du Mésozoique et Tertiaire inférieur pourraient être expliquées d'une manière plus plausible en supposant un champ magnétique variant rapidement, plutôt qu'une dérive continentale à grande échelle, avec ou sans déplacement du pôle de rotation.

3. Les champs magnétiques du Carbonifère et particulièrement du Permien étaient très "stables" en comparaison de ceux des périodes antérieure et postérieure, et avaient une configuration profondément différente de celle du champ actuel.

4. Le champ magnétique précambrien avait une configuration très différente de celle du champ actuel, mais, si l'on tient compte de la durée représentée, était remarquablement constant pour tous les continents.

ZUSAMMENFASSUNG

Diese Zusammenstellung ist ein Versuch, wichtige Informationen zusammen zu bringen und zu diskutieren, die den Magnetismus der Gesteine behandeln, vor allem solcher, die die paläomagnetische Bedeutung haben. Alle den Autoren zur Verfügung stehenden paläomagnetischen Messungen sind hier zusammengestellt und ausgewertet worden, mit einem Schlüssel in Englisch und Russisch für die zusammenfassende Tabelle und die Illustrationen. Die Prinzipien, auf welchen die Auswertung der paläomagnetischen Messungen basiert, sind zusammengefasst, wobei besonderes Gewicht auf statistische Methoden und auf die Ergebnisse und Versuche betreffend magnetischer Stabilität und paläomagnetischer Anwendbarkeit gelegt wird.

Die Auswertung der zusammengestellten Ergebnisse führt zu den folgenden allgemeinen Schlussfolgerungen:

1. Das durchschnittliche magnetische Feld der Erde vom Beginn des Oligozäns bis zur Gegenwart ist fast gleich geblieben, vermutlich infolge eines Dipols im Mittelpunkt der Erde, wobei der Dipol parallel zu der augenblicklichen Rotationsaxe orientiert ist.

2. Die paläomagnetischen Ergebnisse aus dem Mesozoikum und dem Unteren Tertiär können besser durch ein verhältnismässig schnell wechselndes magnetisches Feld als durch eine weiträumige Kontinentalverschiebung, mit oder ohne Wanderung des Rotationspols, erklärt werden.

3. Verglichen mit denen früherer oder späterer Zeiten, waren die magnetischen Felder des Carbons und besonders des Perms, sehr "beständig". Sie waren vollkommen verschieden von der augenblicklichen Gestaltung des Feldes.

4. Das magnetische Feld des Präkambriums war verschieden von der augenblicklichen Konfiguration des Feldes und, wenn man die Zeitspanne in Betracht zieht, auffallend einheitlich für alle Kontinente.

ОБЗОР ЯВЛЕНИЙ ПАЛЕОМАГНЕТИЗМА

Аллан Кокс и Ричард Р. Долл

Резюме

Настоящий обзор является попыткой собрать воедино и обсудить сведения о намагничивании пород, особенно данные, имеющие значение. Все доступные авторам данные о палеомагнетизме были тщательно проанализированы. условные обозначеиян на сводной таблице и на рисунках снабжены пояснениями на английском и русском языке. Сделано обобщение принципов, на которых основана оценка палеомагнитных измерений. Особо подчеркивается значение статистических методов и данных об измерении магнитной стабильности, в свете их применения к изучению палеомагнетизма.

Анализ собранных данных приводит к следующему общему заключению:

1. В течении олигоцена, вплоть до позднечетвертичного времени, среднее магнитное поле земли приближается весьма близко к полю, образованному магнитным диполем в центре земли, ориентированному параллельно к современной оси вращения.

2. Результаты измерений палеомагнитных свойств мезозойских и третичных пород свидетельствуют об относительно быстро изменяющемся магнитном поле, нежели о крупных перемещениях материков, со смещением или без смещения полюсов вращения.

3. Магнитные поля каменноугольного и, особенно, пермского периода были очень "постоянны", по сравнению с полями более раннего или позднего времени, и весьма отличались от современного очертания поля.

4. Докембрийское магнитное поле отличалось от современного очертания магнитного поля и, принимая во внимание промежуток времени, отделяющий нас от докембрия, было в высшей степени постоянным на всех материках.

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INTRODUCTION

Studies of the magnetic properties of rocks have accelerated so rapidly during the past 2 decades that the accumulated information and special techniques of rock magnetism may now well be regarded as a separate geologic discipline. Current interest in the subject has been greatly stimulated by interpretations of the magnetic data as evidence relevant to two persistent geologic hypotheses, continental drift and polar wandering. Moreover, further interest in paleomagnetism has been aroused by observations suggesting that the earth's magnetic field has undergone periodic reversals. While continental drift and polar wandering

hypotheses have been proposed for many decades, and remanent magnetization has been recognized for centuries, a brief description of the recent rise of interest in paleomagnetism through the union of these rather diverse elements will form an appropriate introduction to this review.

The classic early work in paleomagnetism is that of Chevallier (1925), who demonstrated that the remanent magnetizations of several lava flows on Mt. Etna were parallel to the earth's magnetic field measured at nearby observatories at the time the flows erupted. (For earlier studies of rock magnetism *see* the references cited in Chevallier, 1925, and Matuyama, 1929.) Mercanton (1926, p. 860) appears to have been the first to foresee clearly the possibility of using rock magnetism as a tool for testing the theories of polar wandering and continental drift, and he also anticipated the field-reversal hypothesis.

Since these early studies, great progress has been made in understanding the processes by which rocks become magnetized. For the past 20 years, Thellier and his colleagues have been concerned with several of these processes, including magnetization acquired by rocks over long periods of time in weak fields and magnetization by heating and cooling in weak fields. They have applied the results of these studies principally to determinations of the intensity of the earth's field in the past. (*See* Thellier and Thellier, 1959, for an important review of this work.) Important contributions to an understanding of magnetic minerals and magnetizing processes have also been made by other workers, including Rimbert, Haigh, Gorter, Nicholls, and especially the Japanese workers Nagata, Uyeda, Kobayashi, and Akimoto.

The periodic field-reversal hypothesis in its modern form, with the last reversal ending in early Quaternary time, was first proposed by Matuyama (1929, p. 205). Recent interest in reversals stems from Graham's observations (1949, p. 156) of opposing directions of magnetization in sediments, which lead Néel (1951) to propose four mechanisms by which a magnetization might be acquired in a direction opposite to that of the magnetic field acting. Néel's theoretical prediction was brilliantly confirmed when Nagata and Uyeda discovered that the Haruna dacite reproducibly acquired a remanent magnetization opposing the applied field (Nagata and others, 1951; Nagata, 1953b; Uyeda, 1958), suggesting that reversed magnetizations were not due to reversals of the earth's field. However, at about the same time Hospers (1951), working in Iceland, and Roche (1951), working in France, found that the presence or absence of reversals in otherwise indistinguishable lavas depended on the stratigraphic position of the flows. Their data strongly pointed to field reversals, with the most recent one occurring early in the Pleistocene. Research on the reversal problem is continuing, and Uyeda's review (1958) of the self-reversal mechanism of the Haruna dacite is one of the outstanding contributions in recent years.

Methods of using field relationships for demonstrating stability of remanent magnetization have been developed by Graham (1949) and are now standard techniques in paleomagnetic investigations.

With the development of a very sensitive spinner magnetometer by Johnson and McNish (1938), and an astatic magnetometer by Blackett (1952) with a sensitivity close to the theoretical limit, measurements of weakly magnetized sediments became possible. A large number of measurements by Graham (1949; 1955), by Clegg and others (1954a), and by Creer and others (1954) soon confirmed the fragmentary evidence from previous studies indicating that pre-Tertiary rocks do not usually have magnetizations parallel to the present field. Graham (1949) pointed out possible applications to an evaluation of the continental-drift and polar-wandering hypotheses.

Concurrently advances were made by Elsasser and Bullard in explaining the origin of the earth's field by the dynamo theory (reviewed in Elsasser, 1955; 1956). These studies suggested that the earth's axis of rotation and the average axis of the earth's magnetic field should coincide. Encouraged by theoretical developments, Creer and others (1954) proposed a paleomagnetically determined polar wandering path from Precambrian to present times.

When more measurements from other continents became available, it soon appeared to many workers that polar wandering alone would not adequately explain all the data. At the present time relative displacements of nearly all continents, with respect to Europe, have been suggested, for North America by Runcorn (1956b, p. 83) and by Irving (1956a, p. 40), for India by Clegg and others (1956, p. 430), for Australia by Irving and Green (1958, p. 71), for South America by Creer (1958, p. 389), for Africa by Creer and others (1958, p. 500), and for Japan by Nagata and others (1959, p. 382).

One purpose of this review is to bring together the relevant paleomagnetic data. In order

properly to evaluate these data, the reader should be acquainted with the principles of paleomagnetism, and a review of these principles precedes the table of data. After a discussion of these data, we conclude with a short subjective evaluation of some of the paleomagnetic interpretations and suggest a few topics for future study.

Acknowledgments

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The Basis of Paleomagnetism

Magnetization of Rocks

General statement.—The magnetization observed in a rock is determined by two factors, the magnetic field applied to the rock up to the time of observation of the magnetization, and the occurrence of one or more of the several processes by which materials become magnetized. Different magnetizing processes may operate on the rock at different times during its history; the earth's magnetic field may change from time to time, and the magnetization acquired may not in all cases be parallel to the applied field. It is not surprising, therefore, that magnetization is one of the most complex properties that the geologist can study in rocks. Moreover, geologic interpretations of magnetic measurements are critically dependent on the availability of techniques for distinguishing the various magnetic components.

The first two components of magnetization to be distinguished are *remanent* magnetization and *induced* magnetization. Whereas induced magnetization requires the presence of an applied field, remanent magnetization does not. In fields as weak as the earth's, induced magnetization is proportional to the field; the constant of proportionality is called the *susceptibility*. Magnetic-anomaly maps of the earth's field are usually interpreted by assuming that the magnetization in the rocks producing the anomalies is induced magnetization and parallel to the earth's field. However, remanent magnetizations are often not parallel to the earth's field and may be stronger than the induced magnetization. (*See* Nagata, 1953a, p. 128–129, for some typical values.) Therefore, the assumption of a predominance of induced magnetization should be tested by sampling wherever possible.

Each grain of magnetic material in a rock consists of one or more magnetic domains, and although the directions of magnetization are different in different domains, the intensity of magnetization per unit volume, termed the spontaneous magnetization J_S, is the same in all domains of the same mineral. Quantity J_S decreases with increase in temperature and vanishes at the Curie temperature. Sufficiently small grains consist of single domains, and in the absence of an external magnetic field the direction of magnetization in each domain will lie along one of several preferred axes. The directions of these preferred axes and the heights of the magnetic energy barriers that separate them are determined by the shape of the grain, the crystalline anisotropy of the mineral, or both. In an applied magnetic field of increasing intensity the direction of magnetization in the grain is pulled away from the preferred axis toward the field direction. If the energy supplied by the applied field is not greater than the magnetic energy barriers, the direction of magnetization will return to its former position when the field is removed. This magnetization, which is reversible and depends on the applied field, is by definition an induced magnetization.[1]

When the applied field is increased above a critical value termed the *coercive force*, the direction of magnetization in a single-domain grain crosses over a magnetic energy barrier, and when the field is removed the magnetic vector comes to rest along a new direction. In

[1] The term *induced magnetization* is used, as defined here, in geophysical prospecting applications and should not be confused with *induction*, B, of the usual hysteresis curve.

this case the irreversible change in magnetization is a *remanent* or permanent magnetization. Larger grains contain many domains, and, in an increasing magnetic field, domains with magnetizations nearly parallel to the applied field grow at the expense of others. Magnetic energy barriers again prevent an unlimited reversible growth of domains, and the coercive forces of multidomain grains correspond to the magnetic fields necessary to overcome these energy barriers.

A single rock sample may contain several magnetic minerals with a wide range of grain sizes and a wide coercive force spectrum (Graham, 1953, p. 249). The intensity of natural remanent magnetizations is commonly several orders of magnitude less than the maximum intensity that could be developed if the rock were placed in a magnetic field much larger than any of the coercive forces of the different magnetic constituents. This indicates that only a small fraction of the domains in a naturally magnetized rock have a preferred direction of magnetization causing the observed remanent magnetization; the majority have random directions. Whether the domains with preferred orientations occur in constituents with high or low coercive forces depends on the process by which the remanent magnetization was acquired. The natural remanent magnetization acquired by some processes is very "hard" and similar to the remanent magnetization of a good permanent magnet, whereas that acquired by other processes is "soft," corresponding quite closely with the magnetization of "soft" iron. Because rocks are not homogeneous materials and because many of them have been subjected to several magnetizing processes, both types may be found in the same rock. (*See* Graham, 1953, p. 249–252; Clegg and others 1954a, p. 593.) During the past several decades considerable progress has been made in understanding some of the processes by which natural remanent magnetization is acquired by rocks, and in developing techniques for analyzing the observed magnetizations into components corresponding to the various processes.

In the following paragraphs consideration will be given to the principal processes causing natural remanent magnetization in rocks. Processes leading to remanent magnetizations parallel to the applied field will be discussed first. Then factors leading to remanent magnetizations which are not parallel to the applied field will be considered.

Isothermal remanent magnetization.—A rock placed in a magnetic field at room temperature and subsequently removed will acquire a remanent magnetization, provided the field is larger than the lowest coercive force of the magnetic minerals in the rock. The process is simply one in which domains having magnetic energy barriers with corresponding coercive forces less than the applied field align their magnetic moments with the field. When the field is removed, the energy barriers prevent these domains from returning to their former positions, and a net magnetization results. Since minerals in rocks usually have coercive forces of the order of 100 oersted or more, the earth's field of about 0.5 oersted is, in general, not strong enough to produce isothermal remanent magnetization (IRM). On the other hand, the large magnetic fields associated with lightning bolts may impart a substantial IRM to rocks (Cox, 1959). The IRM may easily be removed, or changed in direction, by any field as large as that which produced it.

Viscous magnetization.—If a rock remains in a field too weak to cause IRM for a sufficiently long period of time, it is often possible to measure a new component of remanent magnetization in the direction of the field (*e.g.*, Rimbert, 1956b, p. 2536; Brynjólfsson, 1957, p. 250–251). Such a magnetization, requiring a relatively long time to form, is termed viscous magnetization. In rocks, as in other materials, it is due to the Boltzman distribution of thermal energy which, when converted to magnetic energy, allows the magnetic domains to cross energy barriers that they otherwise could not cross in the weak field of the earth. Although the thermal-energy distribution has a random nature, the weak field of the earth provides a slight bias sufficient to cause a net change of magnetization in the direction of the field. The theory of viscous magnetization is similar to that of thermo-remanent magnetization.

Thermo-remanent magnetization.—Of much more importance for paleomagnetic studies is the process of thermo-remanent magnetization. As a rock cools in the earth's magnetic field it begins to develop spontaneous magnetization at the Curie temperature T_C and a preferential alignment of domains parallel to the field. The resulting magnetization is thermo-remanent magnetization (TRM). It is important to note that not all of the TRM is acquired at the Curie temperature, but rather over a temperature interval extending some tens of degrees below T_C. If, during a cooling experiment, a weak magnetic field is applied only in the temperature interval T_1 to T_2 ($T_2 < T_C$), with zero magnetic field at all other temperatures, a magnetization known as the partial thermo-remanent magnetization (PTRM) is developed.

An example of the PTRM acquired in equal temperature intervals on cooling from the Curie temperature is shown in Figure 1a, where the values are plotted as a function of the mean

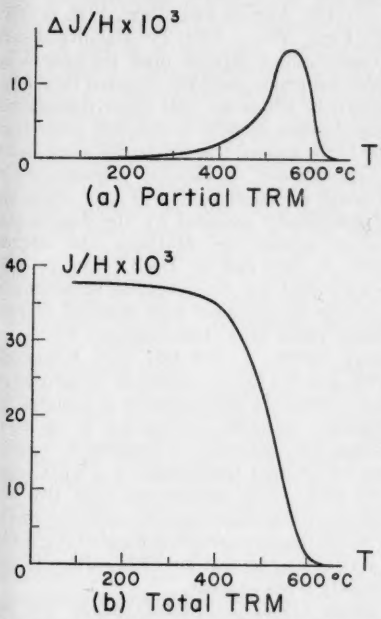


FIGURE 1.—ACQUISITION OF THERMO-REMANENT MAGNETIZATION IN WEAK MAGNETIC FIELDS

(a) Partial thermo-remanent magnetization (PTRM) acquired in field H over equal temperature intervals as a function of mean temperature of interval; (b) thermo-remanent magnetization (TRM) acquired on cooling from Curie temperature to any temperature T in field H and from T to ambient temperature in zero field (for weak field the quantity J/H is approximately constant).

temperature of the interval. Experimentally it is found that for lavas and baked sediments the PTRM acquired in a weak field over any temperature interval T_1 to T_2 is independent of the magnetization acquired in adjacent temperature intervals (Thellier, 1951, p. 213; Nagata, 1953a, p. 142–153); Thellier reports that the rock preserves an exact memory of the temperature and field which produced the PTRM (quoted in Néel, 1955, p. 212). If the rock is heated to temperatures up to T_1 no effect on the PTRM acquired between T_1 and T_2 is observed, whereas it is completely destroyed at temperatures above T_2. The total TRM acquired in a given field is very close to the sum of the PTRM's acquired in the same field between the Curie temperature and room temperature. (*Compare* Figs. 1a and b.)

The theory of TRM for single-domain particles (Néel, 1955, p. 209–212) explains many of these characteristics. In Néel's model (which will be followed here) each grain has two directions in which the magnetic vector can lie with minimum magnetic energy in the absence of a magnetic field; these directions are 180° apart and are separated by a magnetic barrier of energy

$$E = vH_C J_S/2 \tag{1}$$

where v is the volume of the grain, H_C the coercive force, and J_S the spontaneous magnetization of the mineral. When E is greater than the thermal energy (kT), where k is Boltzmann's constant and T the temperature, the thermal fluctuations are not able to move the direction of magnetization across the energy barrier. However, for sufficiently small values of v or sufficiently high values of T, the thermal fluctuations can cause the magnetic moment to move across the barrier. Thus, a total remanent magnetization of initial amount J_0 due to the preferential alignment of a large number of identical single domain grains will, after time t, have decayed to the value J_R given by

$$J_R = J_0 \exp(-t/\tau_0) \tag{2}$$

where τ_0 is termed the *relaxation time*. As in other decay processes, one may speak of the "half-life" of thermo-remanent magnetization which has the value 0.693 τ_0. Quantity τ_0 is given by the equation

$$1/\tau_0 = A(v/T)^{1/2} \exp(-vH_C J_S/2kT) = A(v/T)^{1/2} \exp(-\gamma v/T) \tag{3}$$

Quantities A and γ depend on the elastic and magnetic properties of the minerals, and the other quantities are as defined for equation (1).

An important feature of this model for TRM is that a small change in the quantity (v/T) can cause a very large change in τ_0. For example, the physical constants necessary to evaluate A and γ are known for iron (Néel, 1955, p. 211); and values for the quantity (v/T) of 3.2×10^{-21}, 7.0×10^{-21}, and 9.6×10^{-21} correspond respectively to values of 10^{-1} seconds, 10^9 seconds (3.4×10^2 years), and 3.4×10^9 years for τ_0. At room temperature the grain diameters corresponding to these values of (v/T) are roughly 120 Å, 160 Å, and 180 Å. Thus, the direction of magnetization in a grain with a diameter less than 120 Å is easily and quickly changed by the thermal fluctuations, and the application of a weak field h to a number of

such grains causes a net magnetization in the direction of the field. This "equilibrium" magnetization is given (Néel, 1955, p. 211) by the equation

$$J_E = NvJ_S \tanh (vhJ_S/kT) \qquad (4)$$

where N is the number of grains with volume v.

Because of the strong dependence of τ_0 on (v/T) in equation (3), there is a *critical blocking diameter* for a given mineral, dependent only on the temperature; grains with smaller diameters come to equilibrium very quickly with the magnetization indicated in equation (4), while those with substantially larger diameters maintain their original magnetizations over long intervals of time, regardless of the external field. Similarly, there is a *critical blocking temperature* for all grains of the same diameter.

The acquisition of TRM by single-domain grains is very simple in terms of this model. As a rock cools from its Curie temperature, a given grain assumes the equilibrium magnetization, J_E, until the temperature passes through the critical blocking temperature of the grain. As the temperature goes below this critical value, τ_0 for the grain increases rapidly, and the magnetization becomes "frozen" at the equilibrium level. The independence of partial thermo-remanent magnetizations acquired in different temperature ranges is thus explained as due to the magnetization residing in grains of different diameter. This simple theory explains many of the characteristics of TRM such as its great stability to disturbing fields and its remarkably slow decay.

The acquisition of TRM by most rocks is certainly more complex than indicated here, since many rocks contain magnetic minerals differing in physical properties as well as in grain size. Moreover, rocks containing multi-domain grains, and even massive ferromagnetic mineral specimens, also acquire TRM which, commonly, has the characteristics described above. Verhoogen (1959) suggests that the TRM of these materials may reside in small, highly stressed regions within the ferromagnetic crystals.

The Curie temperatures of magnetic materials in igneous rocks lie below 700° C and, in many rocks, below 600° C. The major portion of the natural remanent magnetization measured in many igneous rocks appears to be TRM. (For more complete discussions of TRM *see* Nagata, 1953a, p. 123–192; Néel, 1955, p. 208–218, 225–241; Verhoogen, 1959.)

Depositional magnetization.—As demonstrated in artificially deposited sediments, previously magnetized magnetic particles attain a preferential alignment during deposition and maintain this alignment after consolidation, giving the sediment a remanent magnetization (Nagata and others, 1943, p. 277–279; Johnson and others, 1948, p. 357–360; King, 1955, p. 120). The stability of such a magnetization depends upon the process by which the grains originally acquired their magnetization. (Processes that cause depositional magnetization to have a direction other than that of the applied field will be discussed later.)

Crystallization or chemical magnetization.—Although the magnetization of some sediments is undoubtedly acquired by the depositional process, studies by Martinez and Howell (1956, p. 205) and by Doell (1956, p. 166) indicate that the magnetization of sediments may also be associated with chemical changes taking place after consolidation. Moreover, Haigh (1958, p. 284–285) and Kobayashi (1959, p. 115–116) have shown in the laboratory that a remanent magnetization is acquired by magnetic materials undergoing a chemical change (*e.g.*, reduction of hematite to magnetite) at constant temperature in a weak magnetic field. These authors also show that the stability of this magnetization, under the effects of higher temperature and demagnetizing fields, is very similar to that for TRM, although the intensity is not so great.

Haigh (1958, p. 278–281) points out the theoretical similarity of the processes causing chemical magnetization and TRM of small grains. As the grains of magnetic material grow chemically, the value of the critical quantity (v/T) in the equations for TRM increases because of an increase in v rather than a decrease in T. As the grain grows through the critical blocking diameter appropriate to the temperature at which the chemical reaction occurs, the equilibrium magnetization J_E (equation 4) is, as in the case for TRM, effectively frozen in. Theoretically, the stability properties of crystallization magnetization and TRM should be similar, and laboratory experiments indicate that this is true.

Self-reversed magnetization.—The most striking example of a magnetization acquired in a direction other than that of the field acting during the acquisition of the magnetization is that of self-reversal. In many paleomagnetic studies directions of magnetization fall into two distinct groups nearly or exactly opposed to each other. Two interpretations of this phenomenon have been proposed: that the earth's magnetic field may periodically reverse itself,

or, alternatively, that some rocks may become magnetized in a direction opposite to that of the field acting on them by a process called self-reversed magnetization.

Several mechanisms may theoretically give rise to self-reversed magnetization. The first to be considered requires two magnetic constituents A and B in the rock. Constituent A has a higher Curie temperature than B and acquires a TRM parallel to the applied field. As the rock cools through the Curie temperature of constituent B, the TRM of constituent A acts by one of several interaction mechanisms to order the magnetization of B in a direction exactly opposite to that of A and, hence, reversed with respect to the original applied field. A self-reversal occurs if, after cooling, the total magnetization of B exceeds that of A (Néel, 1951, p. 92), or if constituent A is later selectively removed chemically (Graham, 1953, p. 252–255).

The simplest type of interaction is magnetostatic (Néel, 1951, p. 100; Uyeda, 1958, p. 50–56), in which the field in the region of constituent B at the time the temperature passes through its Curie point is controlled by the magnetization of A, and is reversed with respect to the applied field. The relationship is shown schematically in Figure 2. For this type of interaction to lead to a self-reversal, very stringent requirements are placed on the geometrical arrangement of the two constituents and on the ratio of the applied field to the spontaneous magnetization of constituent A when B becomes magnetized. In rock-forming minerals this mechanism could occur only in very weak applied fields; it is possible but rather improbable in fields as strong as the earth's, and no example has been found in nature (Uyeda, 1958, p. 52).

A second type of interaction between the two constituents is an exchange interaction across their common boundary. If good registry exists between the crystal lattices of the two constituents, the spontaneous magnetizations on one side of the boundary will tend to become aligned either parallel or antiparallel to the spontaneous magnetization on the other side. The Weiss-Heisenberg exchange interaction between spinning electrons, which is also responsible for spontaneous magnetization, provides the coupling, which may be very strong. Uyeda (1958, p. 104) finds that members of the ilmenite-hematite series $xFeTiO_3 \cdot (1 - x) Fe_2O_3$, with $.45 < x < .6$, become self-reversed, even when the applied field is as high as 17,000 oersted. This type of interaction appears to be responsible for the reversed magnetization of the Haruna dacite (Uyeda, 1958, p. 120), which is one of the two or three rocks reported to be reproducibly self-reversing.

The spontaneous magnetization of some

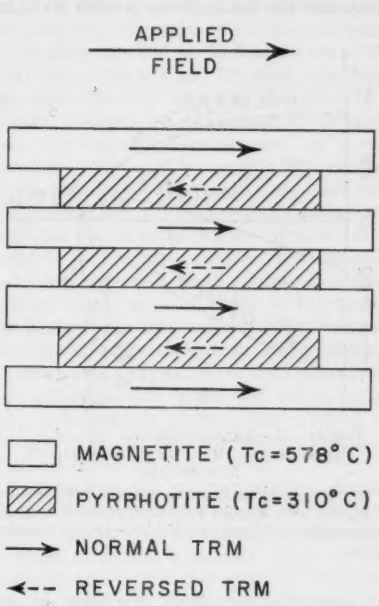


FIGURE 2.—REVERSAL IN PYRRHOTITE CAUSED BY MAGNETOSTATIC INTERACTION BETWEEN TWO DIFFERENT CONSTITUENTS

On cooling, magnetite with higher Curie temperature becomes magnetized first. Further cooling results in magnetization of pyrrhotite in "reversed" field between magnetite layers. Net TRM is 'normal". (After experiment by Uyeda, 1958)

minerals, for example magnetite, is actually made up of two superimposed opposing spontaneous magnetizations, each associated with a separate sublattice in the magnetic mineral. If these two spontaneous magnetizations have different temperature coefficients, the total net spontaneous magnetization may change sign with temperature, as shown in Figure 3. This type of self-reversal mechanism has been demonstrated by Gorter and Schulkes (1953, p. 488) in certain synthetic materials but has not been found in rocks.

A mineral may also undergo self-reversal when cations migrate from disordered to ordered distributions on cooling (Néel, 1955, p. 204; Verhoogen, 1956, p. 208). Moreover, when cooled quickly, cations may be frozen in a disordered state corresponding to a high-temperature equilibrium. Over very long

periods of time the cations will then slowly migrate to the equilibrium-ordered positions, and the process may be accompanied by a self-reversal of the TRM. Verhoogen (1956, p. 208) shows that this mechanism is possible for natural magnetites containing impurities; he estimates that the ordering process would require at least 10^5 to 10^6 years. Such a self-reversal mechanism would therefore not be reproducible in the laboratory.

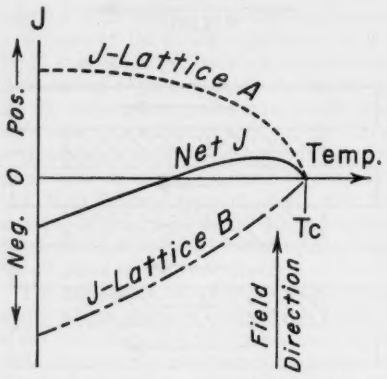


FIGURE 3.—SELF-REVERSAL BY J_S—T DIFFERENCES IN TWO ANTIPARALLEL SUBLATTICES

On cooling, sublattice A is initially dominant and is aligned with applied field. Sublattice B is locked antiparallel to A and is dominant at low temperatures.

This brief and incomplete review of a rather large field of research serves to emphasize several important points about reversely magnetized rocks. The reversed magnetization of some rocks is now known to be due to a self-reversal mechanism. Moreover, many theoretical self-reversal mechanisms have been proposed, and additional mechanisms will doubtless be suggested in the future. However, in order definitely to reject the field-reversal hypothesis it is necessary to show that *all* reversely magnetized rocks are due to self-reversal. This would be a very difficult task since some of the self-reversal mechanisms are difficult to detect and are not reproducible in the laboratory. A further discussion of this problem will be postponed until some of the relevant paleomagnetic data have been considered.

Other processes affecting remanent magnetization.—King (1955, p. 120), in his experiments on artificially deposited varved silts, found that the inclination measured in the samples ranged some 20 to 30 degrees less than the inclination of the field acting, although the declination was faithfully reproduced. This "inclination error" decreased as the field inclination approached the vertical or horizontal. Like most sedimentary minerals, magnetic mineral grains are rarely uniformly equidimensional; moreover, the common magnetic minerals tend to have directions of magnetization parallel to their longest dimension. The "inclination error" arises during the depositional process since the grains will tend to lie with their longest dimension, and hence magnetic direction, parallel to the horizontal bedding plane and not exactly along the applied field direction. King has also demonstrated that an error in the direction of magnetization can occur due to rolling of grains as they settle on the bottom, caused either by deposition on sloping surfaces or by currents.

The magnetic properties of minerals resemble other physical properties in that they are not, in general, completely isotropic; in particular, individual mineral grains usually cannot be magnetized with equal ease in all directions. In all minerals there exist easy and hard directions of magnetization systematically oriented with respect to the crystal lattice, a property called *magneto-crystalline anisotropy*. A single crystal of magnetite, for example, is magnetized more easily along the [111] axes than along the [100] axes, and a crystal of hematite much more easily in the *c* plane than along the *c* axis. A second factor causing anisotropy is the shape of the individual grain. An aggregate of randomly oriented magnetite crystals should have no crystalline anisotropy, but a single grain of the aggregate will be more easily magnetized parallel to its longest dimension. In any of the magnetization processes considered above, except depositional magnetization, in which the grains are already magnetized, the magnetization direction of a single crystal or of an elongated grain will lie between a direction of easy magnetization and the direction of the applied field. However, when preferred-crystal directions or longest-grain dimensions are randomly oriented within a rock sample, the net magnetization direction will be that of the applied field.

Deformation of rocks with a remanent magnetization may also cause a change in the magnetization due to a mechanical rotation of the magnetic particles. A vertical compaction in sediments might, for example, be expected to reduce the inclination of the magnetization

vector (Clegg and others, 1954a, p. 596). Graham (1949, p. 156–158) has considered the effects of plastic deformation on remanent magnetization in the limbs of a fold. However, this phenomenon has rarely been cited as a cause of scattered directions of magnetization, probably because highly deformed beds are usually not chosen for paleomagnetic investigations.

Magnetostriction—the effect of stress on magnetization—is another phenomenon which may be important in the magnetization of rocks. In the investigation by Graham and others (1957, p. 471–472) axial compressive stresses of slightly more than 2500 lbs/sq. in. changed the magnetization in the rocks studied (mostly gneisses and iron ores) by as much as 25 per cent; moreover, the magnetization of some of the samples did not return to the original state after the stress was removed. Many rocks are subjected to large stresses during their histories—the stresses developed during the cooling of basalt, for example, are sufficient to fracture the rock—, and the research described above strongly suggested that magnetostrictive effects might, in general, cause the recorded remanent magnetizations of rocks to be in directions that are not those of the fields acting when remanent magnetization was originally acquired. Stott and Stacey (1959, p. 385) investigated this possibility for TRM by cooling several types of igneous rocks (including basalts, dolerites, andesites, and rhyolites) from above their Curie temperatures in the earth's field while under compressive stresses of 5000 lbs/sq. in. Identical samples were similarly cooled without an applied stress, and in all cases the resulting TRM, measured at room temperature after the stress had been removed, was parallel to the applied field.

Since some magnetostrictive processes may be time-dependent (Graham and others, 1959), field tests are also of interest in evaluating the role of magnetostriction in paleomagnetism. Different magnetic minerals respond in different ways to the same stresses; thus the consistency of results from rocks of the same period that have different mineral assemblages, or were magnetized by different processes, or have had different stress histories would indicate that, for such rocks, magnetostrictive effects have not been important.

Tests for Paleomagnetic Applicability

General statement.—When a study of the remanent magnetism of a suite of samples from a given geologic formation is undertaken, the paleomagnetist is usually less interested in the magnetism itself than in the direction of the magnetic field that produced it. A paleomagnetic study of rocks should therefore yield two pieces of information: the average direction of the magnetic field at the locality where the rocks were collected, and the time or geologic age when the field had that direction. It is usually assumed for paleomagnetic purposes that the magnetization measured is in the direction of the earth's magnetic field existing at the time rocks were magnetized, and that the magnetization was acquired during the formation of the rocks or soon after. We have noted in the preceding sections, however, that rocks may receive a magnetization in several different ways, some of which do not satisfy the assumptions just outlined. For example, depositional magnetization or TRM acquired by rocks with a crystalline or shape anisotropy may not be parallel to the field acting during the magnetization process; and viscous or chemical magnetization may be acquired long after the formation of the rocks.

Fortunately, the magnetizations acquired by the different processes commonly have very different properties which in many cases can be investigated in the laboratory. Many of the magnetic anisotropic properties of rocks can also be measured. Finally, certain geological field tests give very definite limits to the time at which the magnetization took place. The importance of these tests in paleomagnetic studies cannot be overemphasized. Because the critical reader must know whether or not the magnetization was acquired at the time of formation of the rocks, and also whether or not it was acquired parallel to the field acting, it is important to consider the field and laboratory tests in some detail.

Field tests.—Consistency among the directions of magnetization of many samples is sometimes used as a criterion for stability. Although this test is far from conclusive, directions of magnetization that are tightly grouped away from the present field direction have more significance than is often realized. Such a consistency demonstrates immediately the absence of a dominant component of magnetization parallel to the present field, such as might be caused by viscous magnetization or chemical magnetization associated with surface weathering. Moreover, gross petrologic differences of rocks within a formation are usually recognized, and similar differences exist in the magnetic minerals. These differences are commonly indi-

cated by large differences in the intensity of magnetization from sample to sample. Consistency of directions of magnetization in such

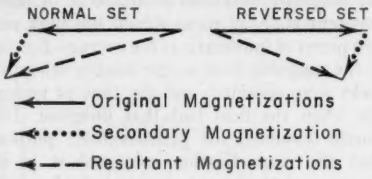


FIGURE 4.—CONSISTENCY-OF-REVERSALS TEST FOR STABILITY

Two sets of magnetization with initial direction 180° apart are no longer exactly reversed if secondary component has been added.

reversals. This test applies to reversals due either to field or self-reversal, since in both cases the mean directions of magnetization are 180° apart. If, subsequent to the original magnetization, both groups acquire an additional component of magnetization as shown in Figure 4, the two resultant groups will no longer be 180° apart. This test is very powerful, since it is also valid for completely homogeneous groups of samples and does not depend on the relative intensity or direction of the secondary magnetization.

In the above field tests the tacit assumption has been made that the rocks have not been tilted or folded. Rocks are, of course, subjected to folding, and Graham's classic fold

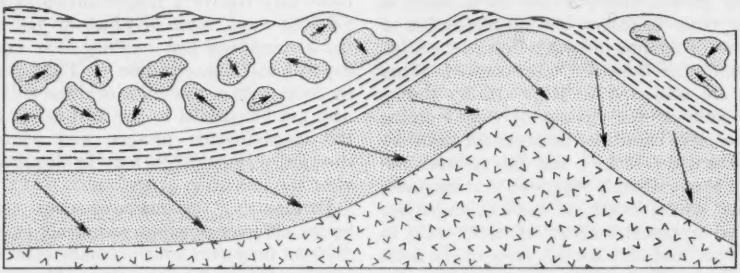


FIGURE 5.—FIELD RELATIONSHIPS INDICATING STABLE MAGNETIZATIONS BY GRAHAM'S "FOLD" AND "CONGLOMERATE" TESTS

a case strongly suggests that the magnetization was acquired in an unchanging magnetic field. If it were acquired by one or more processes acting when the field had different directions, the inhomogeneities would probably result in magnetization directions spread out between the two field directions rather than tightly grouped at some fixed angle between them. Magnetic directions of samples from the same formation are frequently distributed along a plane passing through the present direction of the earth's field (Runcorn, 1956a, p. 305; Creer, 1957a, p. 132–136; Howell and Martinez, 1957, p. 390). The consistency test is not satisfied in this case, and two components of magnetization in varying amounts are present, one of which is parallel to the present field. The significance of a consistency test depends largely on the extent of the sampling and the range in size and composition of the magnetic minerals represented.

Parallelism between tightly grouped mean directions of magnetization in two groups of samples which are reversely magnetized with respect to each other is a much stronger test than simple consistency of directions without

test (Graham, 1949, p. 158) uses folding to establish stability of magnetization. The test is very simple and has great significance. Suppose the directions of magnetization of samples collected from one limb of a fold have a mean direction significantly different from the mean direction of samples collected from the other limb (see Fig. 5). If on conceptually "unfolding" the beds and rotating the directions of magnetization along with them, the mean directions from the two limbs coincide, then the following conclusion is valid: the beds received a magnetization of uniform direction at some time prior to the folding, and the magnetization has not subsequently changed direction. The application of this test to some Precambrian sedimentary rocks is shown in Figure 6. This "tilt correction" is usually made by rotating the beds into the horizontal about the strike direction, a procedure which tacitly assumes that the axes of the folds are horizontal. If the fold is plunging and the magnetic inclinations are other than vertical, this method of correction can lead to serious errors. An extreme example showing how a serious error may be introduced is shown in

Figure 7. The field direction erroneously reconstructed by the simple "tilt correction" differs in azimuth by 90° from the correct direction, and, moreover, a false "reversal" has been generated. Although errors this large occur only for steeply plunging folds and small inclinations, one should, before applying the simple tilt correction, be sure that the fold axes are horizontal. A proper correction for plunging folds can, of course, be made with an additional operation.

The conglomerate test of Graham (1949, p. 158) may also be used to establish magnetic stability. The stability of a formation is tested by locating cobbles or pebbles from the formation in stratigraphically higher conglomerates and measuring the directions of magnetization in these fragments. Since aligning forces associated with the magnetic moment of these large fragments are very much smaller than other forces acting during deposition, the earth's field will not be effective in aligning them. Therefore, a completely random set of directions from the fragments is to be expected if the fragments are stably magnetized. Stability of magnetization of the parent formation is then usually inferred from random directions of magnetization in the fragments, as depicted in Figure 5. Care must be taken in establishing that the random magnetization of the fragments

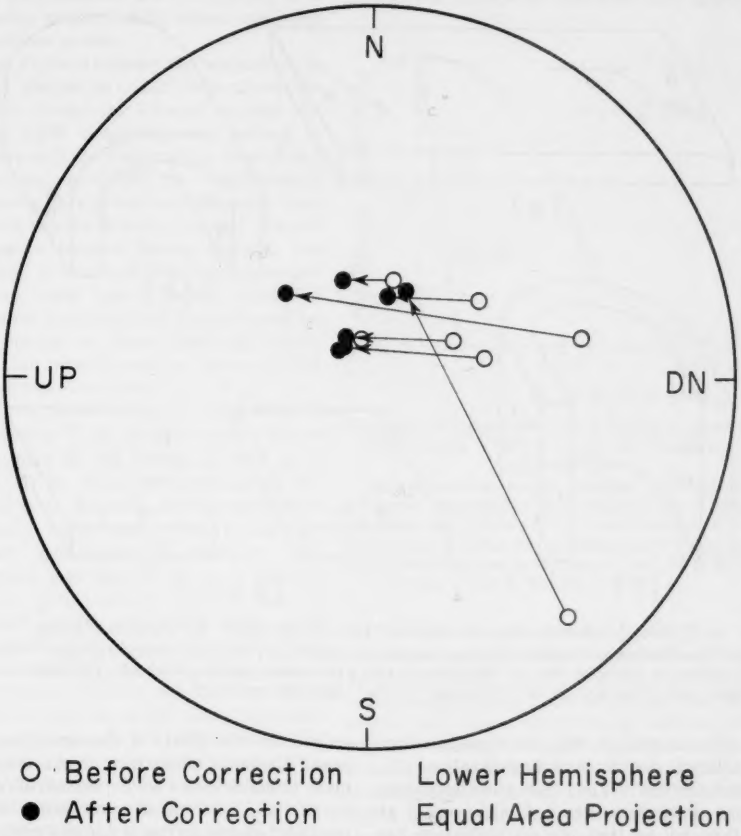


FIGURE 6.—REDUCTION IN SCATTER OF DIRECTIONS OF MAGNETIZATION BY APPLYING CORRECTION FOR TILT OF BEDS

Graham's fold test applied to directions of magnetization in folded Keweenawan sediment. (After Du Bois, 1957)

has not been caused by other than the depositional process, and the test gains in significance when different samples from the same fragment have parallel magnetizations, while samples from adjacent pebbles with the same lithology have different directions of magnetization.

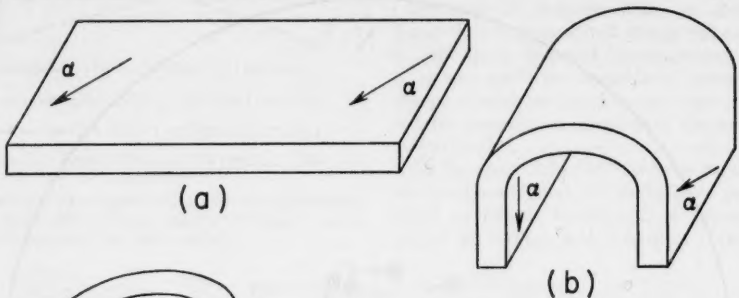


FIGURE 7.—APPLICATION OF SIMPLE "TILT CORRECTION" TO PLUNGING FOLDS

(a) Original uniform directions of magnetization; (b) direction in fold with horizontal axis—"tilt correction" restores to condition (a); (c) directions in fold with steeply plunging fold axis; (d) result of applying simple "tilt correction" to fold (c)—false "reversal" has been generated.

Laboratory tests.—Field tests yield important but not particularly detailed information; at best they tell us that the magnetization has been stable since the occurrence of some event such as folding. Laboratory tests, on the other hand, give more specific and detailed information useful in unraveling the often complex nature of the magnetization found in rocks. Laboratory techniques are also useful for "washing" out unstable components of magnetization as well as the effects of other randomizing processes. Much is now known about the properties of some types of magnetization due, in large part, to the extensive and careful experiments of Nagata and his group, and to the works of Thellier, Rimbert, and Haigh. Thus, it is now often possible, by laboratory analysis, to distinguish the principal components of natural magnetization found in rocks.

An important laboratory experiment is that of examining the magnetic properties of a rock under the effects of demagnetizing processes. Magnetic minerals in rocks consist of many domains with a wide spectrum of coercive forces, and, as noted previously, natural remanent magnetization is due to a preferential alignment of only a few per cent of these domains. Different magnetizing processes tend selectively to align domains concentrated in different parts of the coercive force spectrum, and by means of demagnetization techniques it is possible to learn whether a given natural remanent magnetization resides in domains with low coercive forces ("soft" magnetization), high coercive forces ("hard" magnetization), or perhaps is distributed throughout the coercive force spectrum. In a demagnetization analysis, the "soft" magnetization in the rock is destroyed first by giving low coercive force domains a random orientation; the remaining remanent magnetization is then meas-

ured, and the process is repeated with progressively stronger demagnetizations. Two demagnetization processes may be used: heating the rock to a given temperature followed by cooling in zero magnetic field, or placing it in an alternating magnetic field, whose amplitude slowly decreases to zero.

Although Figure 1b shows the acquisition of TRM as a sample is cooled from above the Curie point, it may also be used to show the amount of TRM remaining after heating to any temperature. As discussed in more detail in the section on TRM for single-domain grains, heating to a given temperature in zero field causes a random orientation in all domains with magnetic barriers having energies less than or equal to the thermal energy. An upper temperature limit beyond which heat demagnetization is not useful is frequently set by chemical changes or phase transitions which may occur at temperatures as low as a few hundred degrees Centigrade.

If a rock is placed in an AC magnetic field with peak value $\tilde{H}$, all domains with coercive forces less than $\tilde{H}$ will follow the field as it alternates. As the AC field is then slowly decreased to zero, domains with progressively lower coercive forces become fixed in different orientations, and hence all domains with coercive forces less than $\tilde{H}$ will have random orientations. If a constant magnetic field is superimposed on the alternating field, or if the variation of the magnetic field with time is not symmetrical, an anhysteretic magnetization will develop (Thellier and Rimbert, 1954, p. 1400) which may mask the remaining remanent magnetization. The development of this magnetization may be prevented by performing the AC demagnetization in the absence of a constant field with the even harmonics filtered out from the current supplying the AC field coil (As and Zijderveld, 1958, p. 310), or by spinning the sample as the alternating field decreases (Brynjólfsson, 1957, p. 248; Cox, 1959, p. 122).

Many of the processes causing remanent magnetization can be reproduced in the laboratory, and demagnetization experiments on such magnetization of known origin are important in interpreting similar experiments on natural remanent magnetism. Figure 8 shows the results of alternating field demagnetization experiments on thermoremanent and chemical magnetizations produced in weak fields and on isothermal remanent magnetization produced in a relatively strong field (Kobayashi, 1959, p. 104). The IRM acquired in a 100-oersted field is effectively destroyed in an alternating field with a peak value of 100 oersteds; however, the TRM acquired in a field of 0.5 oersted has decreased only slightly

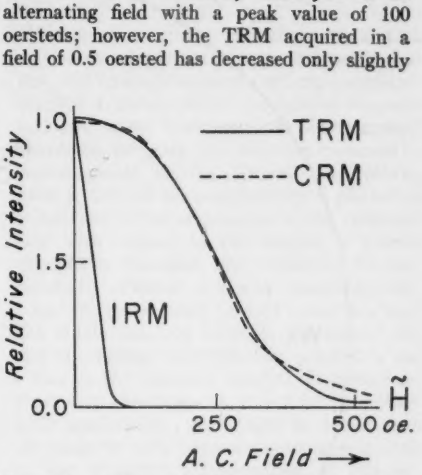
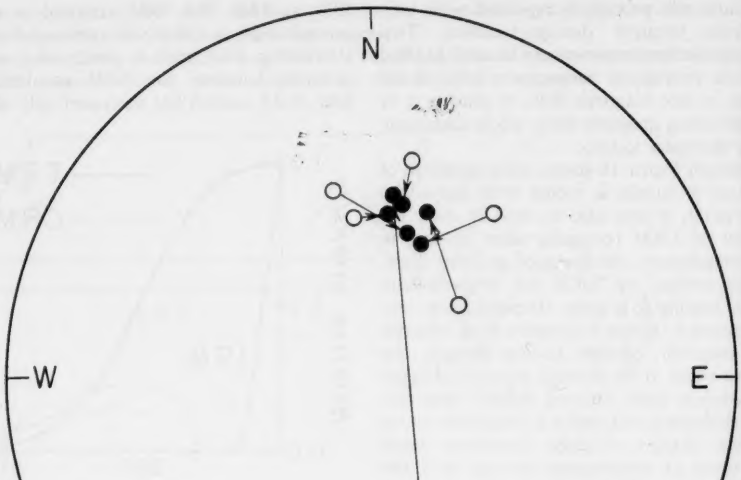


FIGURE 8.—ALTERNATING FIELD-DEMAGNETIZATION CURVES FOR VARIOUS TYPES OF REMANENT MAGNETIZATION

Normalized isothermal, chemical, and thermoremanent magnetizations remaining after demagnetization in A. C. fields, shown as a function of the peak value $\tilde{H}$ of the demagnetizing field. TRM and CRM were acquired in 0.5 oersted field, IRM in 30 oersted field. (After Kobayshi, 1959)

in the 100-oersted alternating field, and a measurable part still remains above 500 oersted. Rimbert (1956a), p. 892 in other experiments noted an appreciable TRM remaining above 900 oersted and only a small change between 500 and 900 oersted. Chemical magnetization has a stability comparable with that of TRM, as was suggested by the similarity of the TRM and CRM theories for single domains. Thus, with these and similar experiments (*see* aspecially Thellier and Rimbert, 1955, p. 1406), it is relatively simple to distinguish IRM in rocks from CRM or TRM, but not to distinguish CRM from TRM.

Viscous magnetization differs from IRM in requiring, for its destruction, an alternating field larger than the field in which it was produced. Rimbert (1956b, p. 2538) found that the magnitude of the AC field needed to destroy viscous magnetizations acquired by volcanic rocks over periods of time up to 2 months varies linearly with the logarithm of the time. For example, the viscous magnetization acquired in a 5-oersted field during 5 minutes required a 37-oersted alternating field for its



○ Original Measurement
● After 300 oe. A. C. Treatment

Lower Hemisphere
Equal Area Projection

FIGURE 9.—DIRECTIONS OF MAGNETIZATION BEFORE AND AFTER ALTERNATING-FIELD PARTIAL DEMAGNETIZATION

All samples are from the same lava flow. (Data from Cox, 1959)

destruction, and that acquired in 2 months in the same field required a field of 180 oersted. Although it is dangerous to extrapolate these results to geologic times, they suggest that viscous magnetization acquired during a million years in a field of 1 oersted would probably be destroyed in alternating fields of the order of a few hundred oersted. A rough verification of the extrapolation may be found in demagnetization studies by Brynjólfsson (1957, p. 251) and Cox (1959, p. 129) in which a viscous magnetization in volcanic rocks about half a million years old was destroyed in alternating fields of 50 to 100 oersted.

Since viscous magnetization acquired in the earth's field and isothermal remanent magnetization due to lightning are probably common sources of scatter in paleomagnetic measurements, these experiments suggest an obvious way of "washing" away these unstable secondary components. Partial thermal and alternating field demagnetization have been used by a number of workers for this purpose (Doell, 1956, p. 165; Cox, 1959, p. 122; Brynjólfsson, 1957, p. 253; Hood, 1958; Creer, 1958, p. 379; As and Zijderveld, 1958, p. 318). Figure 9 shows an example of the effects of alternating field demagnetization on volcanic rocks; most of the initial scatter in these measurements has been shown to be due to lightning (Cox, 1959, p. 135).

Demagnetization experiments are important

in paleomagnetic studies not only for decreasing scatter in the data but also for shedding light on the origin of the remanent magnetization. Moreover, natural magnetizations remaining after demagnetization in fields of the order of 400 oersted are very stable "hard" magnetizations and will certainly not have been disturbed by the effects of sampling, transporting, coring, or measuring operations, or by any process capable of magnetizing only low coercive force domains.

A special series of laboratory tests has been devised by Nagata and hi group (Nagatas and others, 1954, p. 184–185; Nagata and others, 1957, p. 32) for determining whether reversely magnetized rocks represent field or self-reversals. The tests are primarily concerned with the detection of self-reversal properties in the rocks, and for details the reader is referred to the works cited as well as to that of Uyeda (1958).

The field and laboratory tests discussed above are primarily concerned with establishing the stability of natural remanent magnetizations and removing the scattering effects of "soft" magnetizations. However, the very important question of whether the magnetization was acquired parallel to the magnetic field that produced it remains unanswered. In order to devise tests to answer this question one must first consider processes whereby magnetizations are acquired in directions that are not parallel to the applied field direction.

Nonparallel magnetization will be acquired if the magnetic grains in a rock have a shape or crystal anisotropy; rocks with thin layers of magnetite crystals or hematite crystals with parallel axes would possess, respectively, these two types of magnetic anisotropy. Inclination errors associated with depositional magnetization also cause nonparallel magnetization and probably cause anisotropy as well, since flat or elongated grains tend to lie with their longest axes in the bedding plane. Nonparallelism in depositional magnetization may also arise when grains are rolled down inclined depositional planes or moved by bottom currents; Granar (1959, p. 32) has shown that anisotropy will probably be associated with bottom currents, since elongated grains tend to roll with their long axis normal to the current direction.

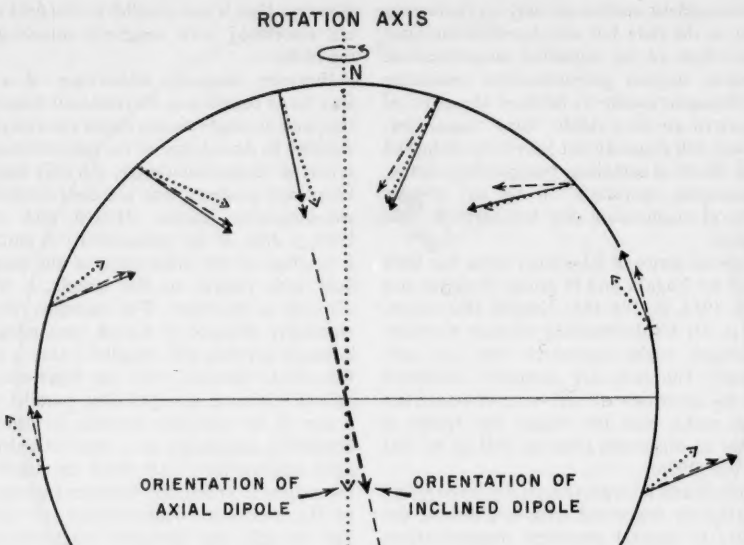
Magnetostriction might also cause a nonparallel magnetization, but tests for its occurrence cannot be devised until the process is better understood. It appears therefore that most processes known to cause a magnetization direction that is not parallel to the field acting are associated with magnetic anisotropy in the rocks.

However, magnetic anisotropy of a rock may be as complex as its remanent magnetization, and no single measurement can completely describe it. Anisotropy of the induced magnetization is, to our knowledge, the only magnetic anisotropy property that has been measured in paleomagnetic studies (Howell and others, 1958, p. 286). If the susceptibility is plotted as a function of the orientation of the magnetic field with respect to the sample, a triaxial ellipsoid is described. For example, the susceptibility ellipsoid of a rock containing only hematite crystals with parallel *c* axes is a very flat oblate spheroid with its short axis, the axis of minimum susceptibility, parallel to the *c* axes of the hematite crystals. In using susceptibility anisotropy as a test for paleomagnetic applicability, care must be taken that the magnetic anisotropy measured corresponds to the remanent magnetization of interest. For example, the remanent magnetization in a rock might be due to hematite with strong susceptibility anisotropy, but this would not be detected if a small proportion of isotropic magnetite were also present.

The Earth's Magnetic Field

Description of field.—The present shape of the earth's magnetic field and its changes during the last several hundred years are of primary importance in paleomagnetism, since these data furnish an estimate of the irregularities and variations likely to be encountered in studies of past magnetic fields. This might be called the expected "signal to noise ratio" for paleomagnetic studies. The present field at the surface of the earth may be described in terms of three components: a relatively small component due to processes occurring above the earth's surface; a dipole component equivalent to the field of a magnetic dipole located at the center of the earth and inclined 11½° from the axis of rotation; and a nondipole component, which would remain if the externally produced field and dipole field were removed.

If the earth's magnetic field is represented by means of spherical harmonics, one may easily recognize and separate these three components. The first such analysis was made by Gauss in 1839 and has been repeated at various intervals since (Chapman and Bartels, 1940, p. 639). The results of the analysis are expressed as a series of terms, each a simple algebraic combi-



0.6 GAUSS

Theoretical field due to geocentric axial dipole

Theoretical field due to geocentric inclined dipole

Earth's field, 1945, projected onto meridional plane 290° east

FIGURE 10.—THEORETICAL MAGNETIC FIELDS OF A GEOCENTRIC AXIAL DIPOLE AND A GEOCENTRIC INCLINED DIPOLE, WITH OBSERVED FIELD DIRECTIONS

Plane of projection passes through geomagnetic poles, and observed field is projected onto this plane. Observation points are at 30-degree intervals from geomagnetic pole.

nation of sin $m\phi$, cos $m\phi$, $P_n^m(\cos\theta)$, and appropriate constants, where ϕ is the longitude, θ the latitude, m and n are integers, and $P_n^m(\cos\theta)$ are associated spherical functions. Moreover, processes occurring above the earth are represented by terms that are mathematically distinguishable from those corresponding to processes occurring within the earth. The externally produced field, physically generated by movement of electrical charges in the ionosphere, fluctuates because of atmospheric tidal effects and sunspot activity. Although its magnitude during magnetic storms may exceed several per cent of the total field, the algebraic average is very small.

Of the terms representing the internally

produced field, those for which $n = 1$ are collectively termed the first-order harmonic and those for which $n = 2, 3, \cdots$ are known as the higher-order harmonics. The nondipole component of the earth's field is represented by the higher-order harmonics, and the dipole component is completely described by the first-order harmonic. Therefore, if the earth's magnetic field were due solely to a dipole at the center of the earth, only the first-order harmonic would appear in the analysis, and conversely the first-order harmonic completely specifies the orientation and intensity of a geocentric dipole. Of all the dipoles that might, by various criteria, be chosen best to approximate the irregular field of the earth, the one that is inclined 11½° from the axis of rotation gives the best average fit, in the sense of least squares, over the entire surface of the earth. The point on the surface of the earth toward which this dipole points is, by definition, the geomagnetic pole, and its present co-ordinates are 78½° North Latitude and 69° West Longitude (Finch and Leaton, 1957, p. 316). The closeness of the fit is shown graphically in Figure 10. The present earth's field has been projected into the plane which passes through the geomagnetic pole and the earth's axis of rotation. The directions are shown at intervals of 30° from the geomagnetic pole. For comparison, the field directions caused by an axial dipole are also shown at these points.

Although theoretical considerations and paleomagnetic results suggest that the geomagnetic pole has not always been at its present location, it is important to note that there is no direct evidence that it has moved. It was not until the latter part of the nineteenth century that data adequate for an accurate determination of the geomagnetic pole became available, and since then it has remained within about half a degree of its present location (Bullard and others, 1950, p. 86).

Superimposed on this stable dipole field is the comparatively irregular, rapidly changing nondipole field represented by the higher-order harmonics. The nondipole field is made up of irregularly distributed regions of high and low field intensity which range in diameter from about 25° to 100° (Bullard and others, 1950, p. 70). Moreover, these regions wax and wane much as the centers of cyclonic activity in the atmosphere do, and present rates of change suggest an average life for an individual cell of the order of 100 years (Elsasser, 1956, p. 87). The movement of these nondipole features over the surface of the earth is not entirely random but shows a systematic westward drift, estimated by several methods at one-fifth degree of longitude per year. The movement is independent of the latitude of the feature (Bullard and others, 1950, p. 83).

The geomagnetic pole does not coincide with the magnetic dip pole, which is defined as the place where the horizontal component of the earth's field vanishes, because a horizontal component due to the nondipole field is present at the geomagnetic pole. At the magnetic dip pole, the nondipole horizontal component exactly cancels the horizontal component of the dipole field. Whereas the geomagnetic pole has not changed since adequate measurements were available, the position of the dip pole has changed relatively rapidly with changes in the nondipole component.

The description of the earth's field in terms of spherical harmonics is a purely mathematical procedure and carries no implication that each term (or group of terms) is physically significant in the sense that it corresponds to a separate physical event. However, other evidence shows that the external field (corresponding to certain of the harmonic terms) is physically different.

With respect to the terms of internal origin, the fact that the first term of the spherical harmonic analysis is predominant does not in itself imply that the dipole term has special or separate physical significance. At greater depths within the earth the higher-order terms become larger relative to the first, and at the core–mantle boundary the two are approximately equal (Elsasser, 1956, p. 87). The strongest evidence that the first term may correspond to a process different from the higher-order terms lies in the observation that their rates of movement relative to the surface during the past 75 years have been very different.

With the exception of the important work of Thellier and Thellier (1959) on the intensity of the earth's past magnetic field, most paleomagnetic data give the field direction only. Therefore, of the various methods of comparing the dipole and nondipole field components, the angular departure of the observed field from the field of a dipole is of most direct interest in paleomagnetism. Examples from four observatories of the angular departure of the observed field from that due to the inclined dipole field (and the axial dipole field) are shown in Figure 11 on an equal-area projection. The axial dipole field is that due to a magnetic dipole aligned along the axis of rotation of the earth and is of more interest in paleomagnetism than is the inclined field. The average departure

between the observed field and the *axial* dipole field at the present time is 8.5° in the northern hemisphere and 17.3° in the southern (Cox, 1959, p. 11). The largest departure disclosed by most of the available observatory records is 29°.

to the geomagnetic pole, is given by the 'dipole" formula:

$$\cot p = \tfrac{1}{2} \tan I \qquad (7)$$

Quantities θ' and ϕ' are the latitude and longitude of the geomagnetic pole; θ and ϕ are

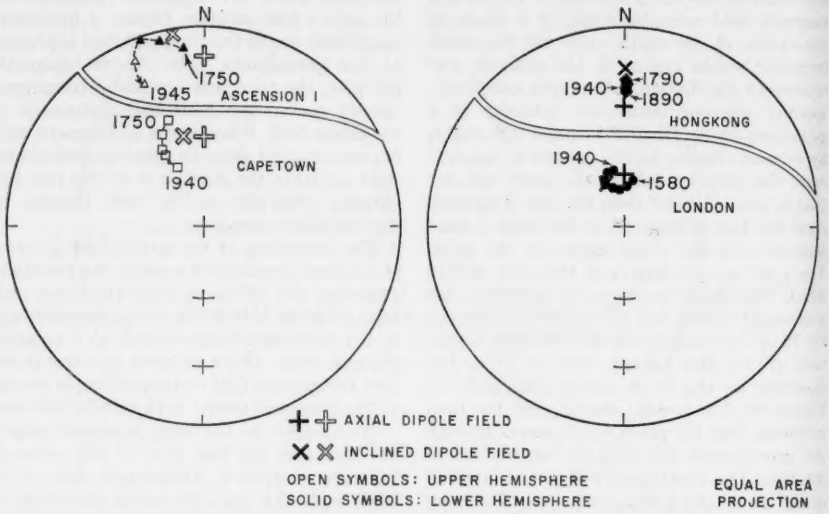


FIGURE 11.—CHANGES IN DIRECTIONS OF EARTH'S MAGNETIC FIELD WITH TIME, AT FOUR MAGNETIC OBSERVATORIES

Successive observations of the field direction are spaced 40–50 years apart. The axial and inclined dipole field directions shown for the four locations are computed from the present positions of the geographic and geomagnetic poles using equations (5) to (7).

It is often convenient in paleomagnetic studies to represent the data not in terms of the field direction measured, but rather in terms of the geocentric dipole that would produce the measured field direction. This is usually done by specifying the geographic co-ordinates of the geomagnetic pole that corresponds to the orientation of this inferred dipole. Given the declination and inclination of the field at an observatory (or as determined paleomagnetically), the position of the geomagnetic pole consistent with the observed direction may be found by the following relations:

$$\sin \theta' = \sin \theta \cos p + \cos \theta \sin p \cos D \qquad (5)$$

$$\sin (\phi' - \phi) = (\sin p \sin D) / \cos \theta' \qquad (6)$$

where p, the angular distance along the great circle from the observatory (or sampling site) the latitude and longitude of the observatory; and D and I are the declination and inclination of the field at the observatory. The position of the geomagnetic pole consistent with a given field direction may also be found graphically by means of a Schmidt or Wulff projection. The relationships are shown graphically in Figure 12.

Equations (5), (6), and (7) establish a one-to-one mapping relation between all possible field directions at an observatory and their equivalent pole locations distributed over the earth's surface—given one quantity, the other is uniquely determined. "Poles" may thus be formally computed from any observed field direction whether due entirely to a geocentric dipole or not. Such poles will here be termed *virtual geomagnetic poles*. When non-dipole components are present, the virtual geomagnetic poles calculated at different

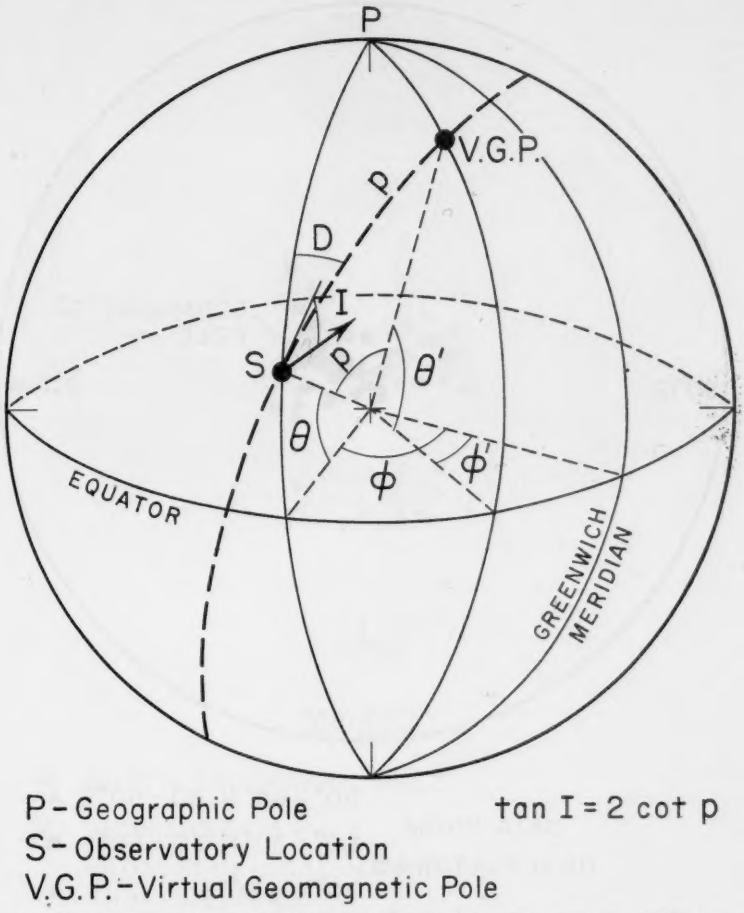


FIGURE 12.—RELATIONSHIPS BETWEEN LOCATION OF OBSERVATORY OR SAMPLING SITE, FIELD DIRECTION, AND VIRTUAL GEOMAGNETIC POLE

Calculation of virtual geomagnetic pole from field-direction data. θ is the latitude and ϕ the longitude of the observatory or site; θ' is the latitude and ϕ' the longitude of the virtual geomagnetic pole; D is the declination, and I the inclination of the field direction; p is the geomagnetic latitude as calculated from equation (7).

localities will not, in general, coincide, and their scatter may be taken as a measure of the departure of the observed field from an ideal dipole field. Figure 13 shows the present-day scatter in virtual geomagnetic poles calculated from the observed field directions at many observatories. Examples of the change in position of virtual geomagnetic poles with time are shown in Figure 14. These poles are calculated from the direction data shown in Figure 11. These figures might therefore suggest an order of magnitude for the "noise signal" to be expected in the determination of average poles by the paleomagnetic method.

Origin of the field.—The problem of the origin of the earth's internally produced magnetic field has long remained one of the least tractable in all of geophysics. Earlier theories suggesting that the field is due to the earth's remanent magnetization do not satisfy two very serious

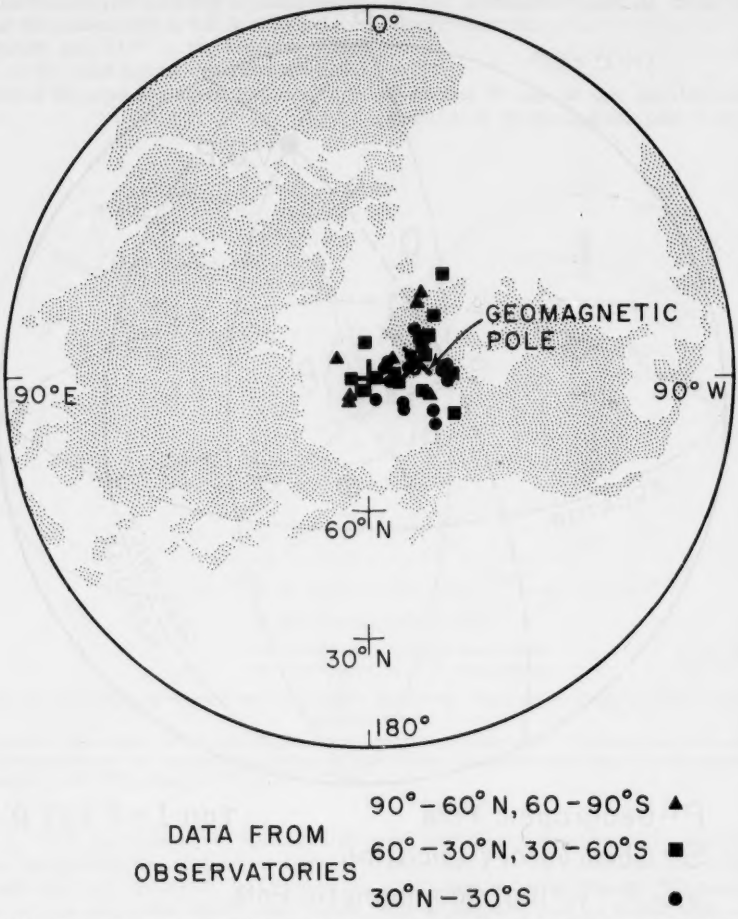


FIGURE 13.—VIRTUAL GEOMAGNETIC POLES CALCULATED FROM OBSERVATIONS AT VARIOUS LATITUDES IN 1945
Data from Vestine and others, 1947

objections. The high rate of change and westward drift of the nondipole field with velocities up to 20 km per year are difficult to explain as due to geologic processes in the mantle or crust. Such processes would certainly proceed at a much more leisurely pace. A further difficulty is that the earth probably does not contain enough ferromagnetic materials below their Curie temperatures to account for the intensity of the field. The Curie temperature of iron is 780° C, of nickel 350° C, and of magnetite 580° C; moreover, there is no evidence that these values increase significantly with pressure. Since the temperature in the earth at depths greater than 25 km is probably above 750° C (Jacobs, 1956, p. 219), it follows that only this outer region could possess a remanent magnetization. The required average intensity would be 6 emu/cc, and in the light of the present knowledge of crustal materials fulfillment of this requirement seems virtually impossible. Magnetite-rich igenous rocks may acquire remanent magnetizations as large as 0.6–1.8 emu/cc in magnetic fields of the order of 1000 oe (Nagata, 1953a, p. 107); however, their natural remanent magnetizations usually range between 0.001 and 0.05 emu/cc.

A theory that most nearly explains all the

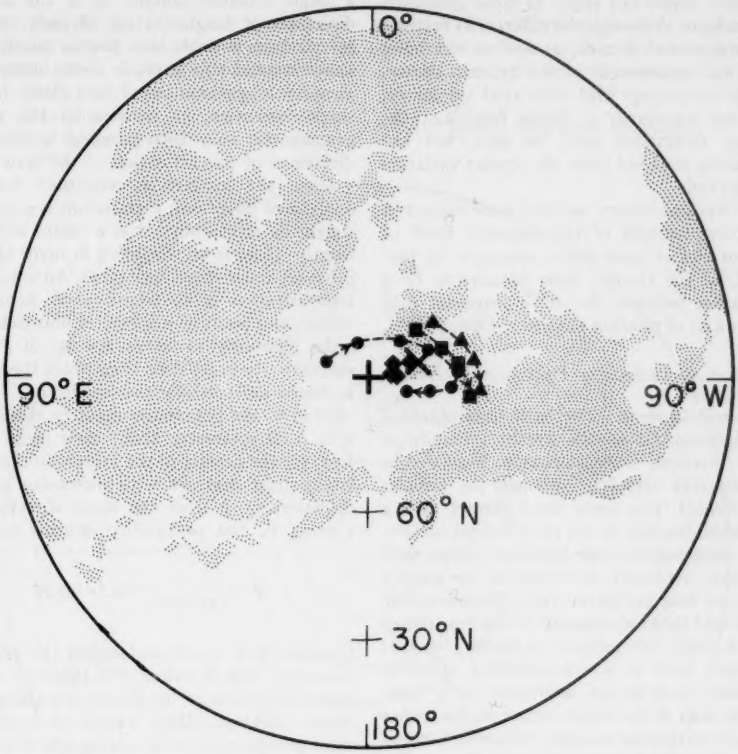


FIGURE 14.—CHANGES IN VIRTUAL GEOMAGNETIC POLES WITH TIME

Poles were calculated from the field-direction data at four observatories shown in Figure 11. Time between points is 40–50 years. (Symbols in Figures 11 and 14 correspond.)

observations is that the fluid core of the earth acts as a self-exciting dynamo (Elsasser, 1956, p. 88–90). In addition to the existence of an electrically conducting fluid the theory requires a source of energy to keep the fluid in convective motion. Moreover, it requires that the fluid be rotating so that order is established in the otherwise random convective motions. The earth's rotation fulfills the last requirement, and aside from transient effects the orientation of the magnetic field should be symmetrically related to the axis of rotation (Runcorn, 1954, p. 61). The rapid changes in the nondipole field are interpreted as the result of fluid eddies near the core–mantle boundary, and the westward drift as due to a smaller angular velocity in the outer layer of the core with respect to the mantle. If we accept this model of the earth's field, then, as a consequence of the relative motion of the core and mantle, the nonaxial components of both the dipole field and nondipole field should cancel when averaged over a sufficiently long time (Runcorn, 1959, p. 91). The dynamo theory requires that one layer in the core be rotating at the same velocity as the mantle, however, and this last

argument would not apply to fields generated in that layer. Although the differential motions of the core and mantle, as well as the basic rotational requirement of the dynamo theory, lead to an average field with axial symmetry, it is not necessarily a dipole field—*i.e.*, the average declination may be zero, but the inclination need not show the dipolar variation with latitude.

The dynamo theory has had some success in explaining features of the magnetic fields of the sun and of some stars; moreover, as predicted by the theory, there appears to be a correlation between the sun's magnetic field and its axis of rotation (Elsasser, 1956, p. 101).

Statistical Analysis of Paleomagnetic Data

General statement.—The basic data obtained in paleomagnetic studies consist generally of many directions of magnetization measured in oriented rock samples. Although the samples are collected from areas never greater than a very small fraction of the earth's total surface, their stratigraphic distribution is often such that they represent directions of the earth's field over long periods of time. Paleomagnetic data might thus be compared with a long record from a single tide gauge in a harbor, whereas the data used in a contemporary spherical harmonic analysis are analogous to a topographic map of the water surface in the harbor at some particular instant. Different sorts of information can be obtained from the two approaches, and different mathematical techniques are necessary for their analyses.

Statistical analysis of sets of vectors or lines.—An analytical tool that has proved very useful for paleomagnetic interpretations is the statistical method developed by Fisher (1953). The method was originally developed for the analysis of paleomagnetic data; however, it is also appropriate for the analysis of other data consisting of sets of vectors or lines. Since Fisher's statistics are of primary importance in the interpretation of paleomagnetic data, and also because they could be applied to other geologic problems, the method will be reviewed in some detail. The analysis of a set of vectors will be considered first, and later the method will be extended to the case for lines.

The statistical method may be used repeatedly on different levels during a single study. For example, an analysis may be made of the directions of magnetization of specimens[2] from a single oriented sample, or of the average directions of magnetization of each oriented sample from a single lava flow or outcrop. An analysis could also be made of the mean directions of magnetization of lava flows from a single formation, or perhaps of the virtual geomagnetic poles corresponding to the mean directions of magnetization of the lava flows.

Since the method is concerned with an analysis of directions, each datum is given unit weight by representing it as a vector with unit length—there is no weighting in favor of more intensely magnetized specimens. An equivalent representation is to regard each datum, or vector, as a point on a sphere of unit radius. In order rigorously to justify the use of Fisher's statistics, the population from which the sample is drawn must satisfy two conditions: (1) the vectors in the population must be distributed with axial symmetry about their mean direction; (2) the density of the vectors in the population must decrease with increasing angular displacement ψ from the mean direction according to the probability density function

$$\mathcal{P} = \frac{\kappa}{4\pi \sinh \kappa} \exp(\kappa \cos \psi) \qquad (8)$$

Quantity κ is a constant called the *precision parameter* and describes the tightness of the group of vectors in the population about their mean direction. High values of κ indicate tight groups, and $\kappa = 0$ corresponds to a population uniformly distributed over the entire surface of the unit sphere. The probability density function $\mathcal{P}$ has the following meaning: given a small area of size δa on the unit sphere, at an angular distance ψ from the mean direction, the proportion of the total points expected in δa is $\mathcal{P}\delta a$. The quantity $(\kappa/4\pi \sinh \kappa)$ is merely a constant factor adjusted so that the integral of $\mathcal{P}$ over the sphere is equal to one. Equation (8) describes a distribution of points on a sphere which is closely analogous with a Gaussian distribution on a plane.

In geologic studies such as petrofabric analyses the density distribution of points on a sphere or hemisphere is often indicated by means of equal-density contour lines. Along each contour line the percentage $\mathfrak{D}$ of the total points included in a circular test area equal to 1 per cent of the area of the sphere or hemisphere

[2] The term *oriented sample* will be used to describe an individually oriented piece of rock; *specimen* will be used to describe pieces of rock cut from an oriented sample and used in the measuring apparatus; *sample* will be used in the usual statistical sense to refer to the data, consisting of a set of N vectors, drawn from a given population of vectors.

remains constant. In Figure 15, contours of equal $\mathcal{P}$ are shown for a Fisher probability distribution with $\kappa = 10$. The density-contour diagram which would result if a large number of vectors from this same population were contoured using the usual 1 per cent test circle is shown in the same figure. As may be seen in the cross sections, the two representations are similar except for a slight leveling of the peak in the density-contour diagram owing to the finite size of the test circle.

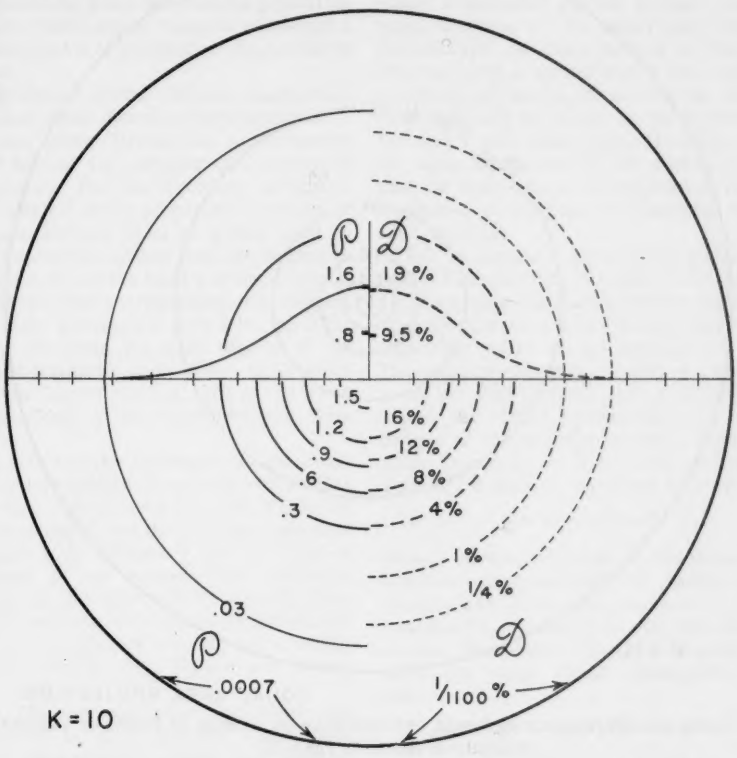


FIGURE 15.—PROBABILITY DENSITY FUNCTION P AND EQUIVALENT POINT-PERCENTAGE CONTOURS
The population of points shown has a symmetrical distribution with mean direction at the pole of the projection (*i.e.*, vertical). $\mathcal{P}$ is the probability density function, $\mathfrak{D}$ is a function showing the distribution of point percentage contours for 1 per cent test areas. Cross sections of these functions are shown along the horizontal line.

Provided the conditions of the statistical model are satisfied, Fisher (1953, p. 296) shows that the direction of the vector sum of the N unit vectors of the sample is the best estimate of the true mean direction of the population. If the ith unit vector has declination or azimuth D_i, east of north, and inclination or plunge I_i below the horizontal, then the mean direction may be calculated from the relations:

$$Z = \sum_{i=1}^{N} \sin I_i \text{ (Downward component)} \tag{9}$$

$$X = \sum_{i=1}^{N} \cos I_i \cos D_i \text{ (North component)} \tag{10}$$

$$Y = \sum_{i=1}^{N} \cos I_i \sin D_i \text{ (East component)} \tag{11}$$

$$R = (X^2 + Y^2 + Z^2)^{1/2} \tag{12}$$

$$\sin I_R = Z/R \tag{13}$$

$$\tan D_R = Y/X \tag{14}$$

where Z, X, and Y are the components of the resultant vector, R is its length, and D_R and I_R its declination and inclination, respectively.

In paleomagnetic analyses, P is usually taken as 0.05, which means that there is 1 chance in 20 that the true mean direction of the popula-

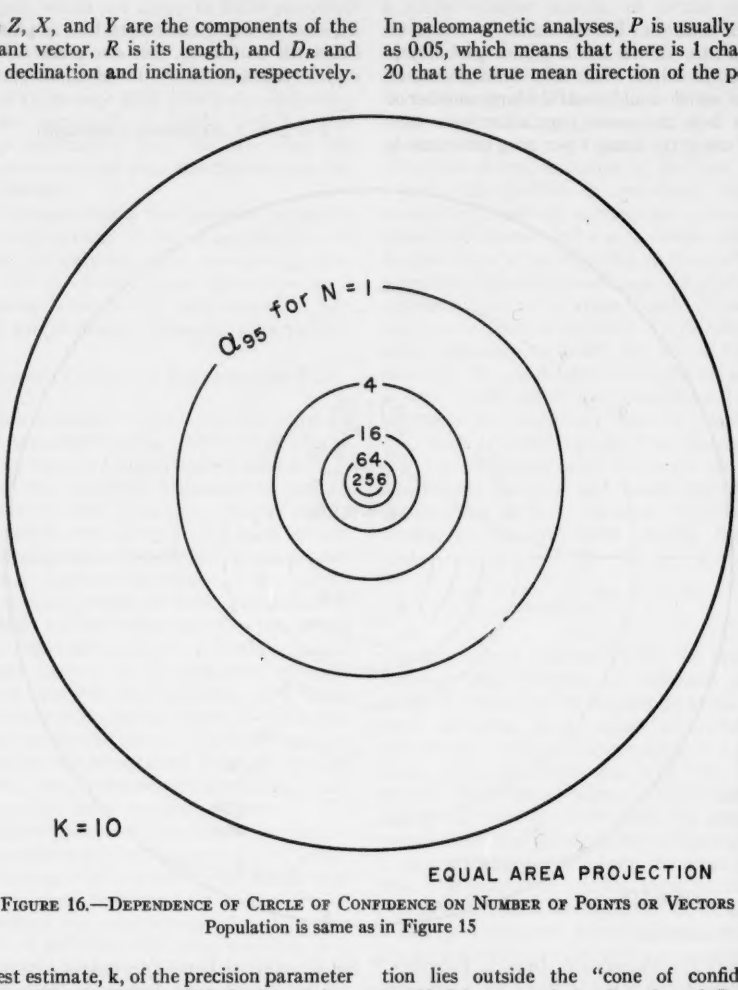


FIGURE 16.—DEPENDENCE OF CIRCLE OF CONFIDENCE ON NUMBER OF POINTS OR VECTORS
Population is same as in Figure 15

The best estimate, k, of the precision parameter κ is given (Fisher, 1953, p. 303), for $k > 3$, by

$$k = \frac{N - 1}{N - R} \tag{15}$$

At a probability level of $(1 - P)$, the true mean direction of the population lies within a circular cone about the resultant vector R with a semivertical angle $\alpha_{(1-P)}$, given (Fisher, 1953, p. 303) for $\kappa > 3$, by

$$\cos \alpha_{(1-P)} = 1 - \frac{N - R}{R}\left\{\left(\frac{1}{P}\right)^{1/N-1} - 1\right\} \tag{16}$$

tion lies outside the "cone of confidence" specified by α_{95} and the direction of R. Some approximate relationships, valid for small values of α, are

$$\alpha_{50} = \frac{67.5°}{\sqrt{kN}} \qquad (P = 0.5) \tag{17}$$

$$\alpha_{95} = \frac{140°}{\sqrt{kN}} \qquad (P = .05). \tag{18}$$

For a discussion of these and other useful relations, reference is made to Watson (1956) and Watson and Irving (1957, p. 289–293).

As the number of vectors included in the

analysis increases without limit, k, the best estimate of the precision parameter, approaches the true value of κ, the precision parameter; on the other hand α becomes infinitely small as N becomes infinitely large. The relationship is shown graphically in Figure 16 for $\kappa = 10$. Thus, even for small values of κ, it is possible to determine the mean direction of a population with any desired degree of accuracy provided a sufficient number of independent measurements are made.

To determine whether the paleomagnetically determined mean direction differs significantly from some known direction such as the present earth's field at the sampling site, α may be used directly. The two directions are significantly different at the probability level used if the angle between them is greater than α. Fisher's statistical method may also be used to calculate α, k, and the mean position of virtual geomagnetic poles corresponding, for example, to the mean directions of lava flows. As is the case for directions, the mean position of the virtual geomagnetic poles differs significantly from some known position, such as the north geographic pole, if the distance between them exceeds α.

It is often desirable to compare one paleomagnetically determined field direction with another rather than with a known direction. A criterion sometimes used is that the two mean directions are significantly different if the two cones of confidence do not intersect, and conversely that they are not different if the cones do intersect. This criterion is not rigorously correct, and more exact significance tests are now available (Watson, 1956, p. 157; Watson and Irving, 1957, p. 293).

Sets of lines or axes rather than vectors are encountered in many geologic applications, and in paleomagnetic studies the problem arises when normal and reversed magnetizations are analyzed together. In this case the field is known to be parallel to a certain line, but the sense along that line is not specified.

If a set of lines or axes has axial symmetry about its mean direction and approximates the density distribution of equation (8), it may be analyzed using Fisher's statistics. However, it is first necessary to convert the lines to vectors—that is, to give an arbitrary sense to each line. This is most easily done by dividing the unit sphere into two hemispheres and regarding each line in one of the hemispheres as a positive unit vector. The vector sum is then calculated in the usual manner. The choice of the plane dividing the sphere is not arbitrary, however; if the vector sum is to be the true mean direction of the lines, it is necessary that it be normal to the dividing plane. In practice the proper plane may be found as follows: (1) a provisional mean direction is estimated and the sphere is divided by the plane normal to this direction, (2) the vector sum of the points in one hemisphere is computed yielding a more accurate mean direction, (3) the sphere may then be redivided by the plane normal to this new direction, and a second vector sum may be computed, (4) finally, the above steps may be reiterated until no change in the direction of the vector sum takes place. Quantity α has the same significance as for vectors, except that the cone defining the confidence interval is reflected in the plane perpendicular to the mean direction.

Ovals of confidence about virtual geomagnetic poles.—Equations (5), (6), and (7) establish a 1-to-1 mapping relationship between any mean field direction at a given locality and a corresponding mean virtual geomagnetic pole. The equations also map the circle of confidence about the field direction into a closed curve around the virtual geomagnetic pole which, because of the mapping function, is an oval rather than a circle. If α is the semivertical angle of the circle of confidence we may write

$$\alpha = dI = dD \cos I \tag{19}$$

where I is the inclination of the mean field direction, and dI and dD are changes in inclination and declination, respectively. From equations (5), (6), and (7) we may then find the semiaxes δp and δm of the oval of confidence about the mean virtual geomagnetic pole from:

$$\delta p = \tfrac{1}{2}(1 + 3\cos^2 p)\, dI = c_1\alpha \tag{20}$$

$$\delta m = \sin p\, dD = \sin p / \cos I = c_2\alpha \tag{21}$$

where p is the distance from the sampling site to the virtual geomagnetic pole (Irving, 1956a, p. 26). The semiaxis δp lies along the great circle passing through the point of observation and the virtual geomagnetic pole, and the semiaxis δm is perpendicular to δp.

The constants c_1 and c_2, which determine the "ellipticity" of the oval of confidence, depend only on the inclination of the field and increase with increasing inclination. Quantity c_1 increases from a value of $\frac{1}{2}$ for 0 inclination to 2 for vertical inclination, and c_2 increases from 1 to 2 over the same range. Thus, for vertical inclinations a circle of confidence about the mean field direction with radius α

maps into another circle about the pole with radius 2α, whereas for flat inclinations the semiaxes of the oval of confidence for the same circle are $\frac{1}{2}\alpha$ and α respectively.

If α is calculated for a probability of 95 per cent, there is a 95 per cent probability that the virtual geomagnetic pole lies within the oval determined from equations (20) and (21).

Sources of error.—Fisher's statistical methods, like others, can be incorrectly applied, and a review of the published data suggests to us that underestimates of α_{95} are not uncommon. Too small a value for α_{95} often results when the N vectors constituting the sample have not been randomly drawn from the same population. This error is analogous to that which would arise if, in finding the mean chemical composition of the Sierra Nevada batholith, five samples were collected at random, 20 separate chemical analyses performed on each sample, and the resultant data treated statistically as if 100 samples had been collected and a single chemical analysis made on each sample.

As a numerical example of the error that may arise in this way, suppose that the true mean directions of magnetization of five lava flows are scattered with a precision parameter κ_F. Let 15 oriented samples be collected from each flow with a between-sample precision parameter κ_{SA}, and let four specimens be measured from each sample with a between-specimen precision parameter κ_{SP}. The scatter in directions of magnetization within samples and within lava flows is usually much smaller than that between flows in the same formation; fairly typical values are $\kappa_F = 30$, $\kappa_{SA} = 200$, and $\kappa_{SP} = 1000$ (Cox, 1959, p. 76).

If each mean sample direction is first found by taking the mean of its four specimens, and if each mean flow direction is then found by a Fisher analysis of its 15 mean sample directions, the resulting estimated mean flow directions will deviate slightly from the true flow directions. Therefore, the estimated precision parameter k_F will be slightly smaller than that for the true mean flow directions, κ_F. However, for the values of κ_{SP}, κ_{SA}, and κ_F indicated, the difference is small, and it can be shown that k_F is equal to 29.7. The circle of confidence using this value is: $\alpha_{95} \cong 140°/\sqrt{29.7 \times 5} = 11\frac{1}{2}°$, by the approximate equation (18). This is a very close approximation to the smallest value of α_{95} that can correctly be calculated from these data.

An incorrect procedure is to analyze the 300 specimen measurements as if each had been chosen randomly from the same population. It can be shown that the "precision parameter" for the 300 vectors is about 25, given the above values for κ_{SP}, κ_{SA}, and κ_F. We may use this value in the approximate formula (18), with N equal to 300, to estimate the "circle of confidence" which would be found in this incorrect analysis; "α_{95}" $= 140°/\sqrt{25 \times 300} = 1\frac{1}{2}°$, which is too small by a factor of 7.

The same problem arises in a paleomagnetic study of sediments, where the different values of κ might describe the precision at different hierarchal levels in the sampling scheme such as the stratigraphic units in a formation, the sampling sites within a stratigraphic unit, or the oriented samples at a site.

When it can be established that a group of data at some level in the sampling scheme corresponds essentially to one point in time, as, for example, the group of measurements from one lava flow, then the variations with time of the earth's field set a lower limit to the value of α_{95} that can be obtained from N groups of data representing N points in time, no matter how many data are in each group. This lower limit for α_{95} may be estimated by the approximate formula (18):

$$\alpha_{95} = 140°/\sqrt{k'N} \qquad (22)$$

where k' is the precision parameter corresponding to variations of the earth's field with time.

A rough estimate of k' may be made by assuming that variations with time of the field at a locality are similar to the present variation of the field around the circle of latitude passing through the locality. Creer (1955) estimates that k' at the equator is about 16 and that it increases poleward, reaching a value of about 70 in high latitudes. Cox (1959, p. 38), using a different method of analysis, finds a similar range of values with large irregularities in the latitude dependence. For the latitudes where most paleomagnetic sampling has been done 30 is a good average value for k' to be used in equation (22). Using this value, a lower limit for the value of α_{95} obtainable from four lava flows is 12°. If the present irregularities in the earth's field existed in the past, some of the reported values of α_{95} based on a large number of samples collected from a few lava flows are probably too small.

One final note should be made concerning the use of Fisher's statistics. The vector population which is represented by the magnetic measurements should satisfy the density distribution given by equation (8) and have symmetry about the mean direction. Watson and Irving (1957, p. 293) have tested these re-

quirements on two sets of stably magnetized rocks and one set of unstable rocks. To the extent allowed by the number of measurements available, the two stable magnetizations satisfied Fisher's requirements, whereas the unstable one did not.

Design of experiments.—The variation in directions of magnetization encountered in a paleomagnetic investigation may arise from many sources, and values of the corresponding precision parameters may differ by several orders of magnitude. Statistical methods are useful in designing sampling schemes and experiments so that the greatest accuracy can be obtained from the smallest possible number of measurements.

Watson and Irving (1957, p. 296) have considered the case of the two-level sampling scheme in some detail. If a total of N oriented samples are collected at B sites or from B lava flows, α_{95} is given by the approximate formula

$$\alpha_{95} \simeq 140^\circ \left(\frac{1}{\omega N} + \frac{1}{\beta B}\right)^{1/2} \qquad (23)$$

where ω is the within-site precision parameter at *all* sites, and β is the between-site precision parameter. Watson and Irving conclude that, whatever the values of ω and β, the smallest number of samples needed to achieve a given α is made up of a single observation at each of the B sites, but that in practice two samples are desirable to test for gross experimental error and magnetic stability. If, as is often the case, the number of sampling sites is limited, a preliminary estimate of ω and β may be used in equation (23) to estimate the number of oriented samples which should be collected at each site. In the example cited by Watson and Irving (1957), the use of this method gave an α_{95} 8 per cent lower than that found using mean-site directions.

Frequently a small α is desired not only for the entire formation but also at each site for possible use in stratigraphic correlation or other geologic applications. From a preliminary estimate of the appropriate precision parameters an estimate of the number of samples required at each site can be made using equation (18).

Other statistical methods.—A measure of the dispersion of a set of vectors that makes no assumption about their density distribution is the angular standard or root mean square deviation $\delta_{r.m.s.}$ defined (Wilson, 1959, p. 755) as the root mean square of the angular distances δ_i of the unit vectors from the mean direction. Wilson (1959, p. 755) shows that

$$\delta_{r.m.s.} = \left\{\sum_{i=1}^{N} \delta_i^2/N\right\}^{1/2} = \{2(N - R)/N\}^{1/2} \qquad (24)$$

where N is the total number of vectors, and R is the length of the vector sum. If the distribution of the vector population is that of Fisher's model, then 63 per cent of the vectors will be within a cone with radius $\delta_{r.m.s.}$ (Creer and others, 1959, p. 316).

Paleomagnetic Data

General Statement

Paleomagnetic results are of interest in several fields of study, and an attempt has been made in Table 1 to assemble all the available basic data in as compressed and accessible a form of reference as feasible. All information available to us that could possibly be used for paleomagnetic purposes is here tabulated. No data have been excluded, even where there is no evidence for stability or where there are other reasons for rejection. It is our belief that sufficient basic data should be available to enable the critical reader to decide for himself whether individual determinations are sufficiently reliable for his purposes. Description of the number of samples collected, their lithology, the areal and stratigraphic extent of the sampling, possible stratigraphic uncertainties, and tests for paleomagnetic applicability are fully as important as a list of pole positions.

The numerical data presented include the coordinates of the sampling locality, the direction of magnetization after correcting for tilt of the strata, and the co-ordinates of the corresponding geomagnetic pole. Confidence limits about the magnetic direction and pole position are also included. In many paleomagnetic studies not all these quantities are given, frequently because the appropriate statistical techniques were not available when the study was made, and in other cases because the original author did not intend that "poles" be calculated from the data. Because of this, separate entries are used for each source of data in Table 1. Extensive use has been made of previous reviews, notably those of Hospers (1955), Creer and others (1954), Runcorn (1955a), Irving (1956a), Creer and others (1957), and Irving (1959). In general, a review is cited only if it lists data which did not appear in the original source or if

there is a difference in the numerical values cited.

The quantities listed in entries designated by an asterisk were found by us in the following ways. The co-ordinates of sampling sites, when not listed in original sources, were located in standard atlases and are reported to the nearest half degree. When the original source lists directions of magnetization with no mean direction or pole position we have made a Fisher statistical analysis of the data, and the results are listed to the nearest half degree. Frequently, directions of magnetization are shown only on stereographic or equal-area projections, and we have scaled the magnetization directions from these diagrams. Since some of the diagrams are small, errors may arise during the scaling process; however, other workers have used this procedure in earlier reviews (*e.g.*, Irving, 1956a, Tables 1, 2, 3; Irving, 1959, Table 1), and it is encouraging to note that the different determinations usually agree to within 1–2 degrees. Several examples showing our values and those of previous reviewers are listed in the tables.

Virtual geomagnetic pole positions can be found from locality co-ordinates and mean directions of magnetization by using equations (5) through (7) or, alternately, by using a stereographic or equal-area projection. The pole positions in Table 1 attributed to the present authors were calculated on a Schmidt equal-area projection 20 cm in diameter. These pole positions were read to the nearest half degree, but because of a slight distortion in the projection used they are probably accurate only to about 1 degree.

Occasionally pole positions but not mean directions of magnetization are listed in original sources. Although it would be possible to recalculate the original direction data from the pole position and sampling-locality co-ordinates using equations (5) to (7), we have preferred to scale the original individual directions of magnetization from diagrams and compute mean field directions from them. The virtual geomagnetic poles calculated from these mean field directions usually differ by only 1–2 degrees from the pole position listed in the original source, again indicating that analyses based on data scaled from small diagrams can be quite accurate. Where this procedure has been employed we have listed the virtual geomagnetic pole corresponding to the mean field direction calculated by us so that the two will be consistent. A pole position calculated by us and differing by 1–2 degrees from that in the original reference is intended as a verification rather than a correction of the earlier result.

In a few cases it has not been possible to reconcile sampling-area co-ordinates, field directions, and pole positions listed in original sources with our computational methods. We have attempted to reach the authors concerned, stating the methods we have used, so that the reasons for the differences might be determined. Where this has not been possible we have used the methods discussed in this paper with the available data so that all results would be as nearly consistent as possible.

The proper application of statistical methods is especially important when paleomagnetic results are interpreted in terms of continental drift and polar wandering, and errors arising from the treatment of each specimen measurement as an independent datum were discussed in the section on statistics. Where adequate data have been given in original sources, and where, in our view, confidence intervals are too low, new analyses have been made for Table 1. In all these cases both the original values and the procedures used by us are described. Even in the absence of complete data, it occasionally has been possible to make realistic estimates of confidence intervals. For example, when specimen measurements have been made on samples from a few lava flows, the secular variation of the earth's field and the number of flows sets a lower limit to α_{95}. (*See* equation (22).) In Table 1 numerous confidence intervals have been recalculated using this equation. For sediments it is usually difficult to judge what level in the sampling scheme represents an independent point in time, and, therefore, few such changes have been made in the statistical data for sediments. However, when the statistical analysis has been applied to many individual specimen measurements made from only a few oriented samples, it is possible that the resulting circle of confidence is too small.

Values of k, the precision parameter describing variations in the observed directions of magnetization, are listed, where possible, for each entry in Table 1 because of the importance of this quantity in evaluating paleomagnetic data. As discussed previously, the present irregularities in the earth's magnetic field in moderate latitudes may be described by a value for k of approximately 30. If this value is taken as a measure of the variations of the field at one locality over an interval of time in the past, then a set of paleomagnetic data in which each datum corresponds to an individual point in

time would be expected to have a similar value of k. Values of k considerably lower than 30 indicate a variation larger than that presently observed in the earth's field and may be due to a greater amount of variation in the past field, to experimental error in measurement, or possibly to the presence of anomalous components of magnetization in the rocks. On the other hand, higher values of k may arise in two different ways. If each specimen measured has effectively averaged the earth's field direction over a long interval of time, then each sample will have a direction close to the mean field direction, and k will be large; alternatively, if many of the samples measured were magnetized at the same time, the mean direction of the group will not be parallel to the average direction of the field, but the scatter will be small, and k will again be large.

The stratigraphic subdivisions of Table 1 are somewhat arbitrary and simply reflect the order in which the data will be discussed. Since many rocks giving consistent paleomagnetic results are poorly dated geologically, the assignment of a particular paleomagnetic investigation to one of the subdivisions in Table 1 has sometimes been difficult. In several instances stratigraphic assignments different from those in previous reviews have been made. In these cases we have noted the original stratigraphic assignment and have, where possible, given some indication of the stratigraphic uncertainty. We feel ourselves unqualified to pursue this problem further, and, for additional information concerning ages, reference is made to the original sources. No stratigraphic ordering is implied by the order of listing in each section of the table.

In evaluating the reliability of individual studies listed in Table 1, the following questions should be carefully considered: (1) Do field and laboratory tests indicate that the magnetization is stable and parallel to the field in which it developed? (2) Do the samples represent enough time to insure that rapid variations in the earth's field have been averaged out? (3) Have appropriate statistical methods been applied, and is the value of α_{95} realistic considering probable variations in the earth's field? (4) Has a sufficiently large geographical area been sampled to insure that the direction of magnetization has not been influenced by local effects such as magnetic field anomalies or small, undetected tectonic movements? (5) Are the geological structure and history sufficiently well known to allow proper bedding corrections to be made and to eliminate the possibility of remagnetization during metamorphism or other alteration? (6) To what geological interval of time and what geographical region do these results apply? (Rather extensive sampling is necessary before a pole position can be regarded as representing, say, "the Carboniferous of Asia").

Key to Table 1

The data are arranged in three sections: the numerical data, a list of references from which these data were obtained, and relevant remarks. Each entry is designated by a serial number and contains data from one source only; there is no significance in the ordering of the entries. The entries in boldface type are those from which the text figures in the next section were made and include values in all columns if these were available or could be calculated from data in the original reference.

Rocks sampled.—The formation name or other descriptive designation from the original reference is given here, together with an indication of what petrologic type is represented. Within each section the results from a given continent are grouped together.

No.—This column contains a number for each entry. In each section numbers begin with 1, and the complete serial designation of an entry is understood to be the letter describing the table subsection followed by the entry number—for example **A 3** and **D 4**.

Locality.—The two columns under this heading give the latitude and longitude of the place where the samples were collected. The latitude (Lat) is given in degrees north (N) or south (S) of the equator, and the longitude (Long) in degrees east (E) or west (W) of Greenwich.

Magnetic direction.—The columns under this heading give the magnetic direction and statistical data. The declination of the average magnetic direction (Decl) is given in degrees east of geographic north, and the inclination of the average magnetic direction (Incl) is given in degrees below the horizontal or above the horizontal (the latter is indicated by a minus sign). Correction has been made for tilt of the strata. The statistical data listed are the circle of confidence in degrees at a probability level of 95 per cent (α_{95}), the precision parameter (k), and the number of vectors (N) used in obtaining these values. Values of k and N apply to magnetic field directions for all those entries which list values for the magnetic direction. In other

entries k and N apply to statistical analyses of sets of pole positions.

Pole position.—These columns give the location of the pole of the theoretical geocentric dipole (the virtual geomagnetic pole) consistent with the co-ordinates of the sampling locality and the mean direction of magnetization. The location is given by the values in columns (Lat) and (Long), with the same conventions as for the sampling locality. Each magnetic dipole has a north and a south pole, and the pole listed is the one that falls in the northern hemisphere. In column (P) which designates the polarity of the pole, the letter S is used if the direction of magnetization corresponds to a magnetic south pole in the northern hemisphere (the present magnetic field is of this type), and the letter N is used if it corresponds to a magnetic north pole in the northern hemisphere. Poles calculated from sets of samples having approximately opposing or reversed polarities are designated (M). (δm) and (δp) are the values in degrees of the semimajor and semiminor axes, respectively, of the 95 per cent confidence oval about the mean pole position, corresponding to the value of α_{95} about the mean magnetic direction in each entry. Quantity δp is measured along the great circle passing through the sampling site and the mean pole position, and δm along the great circle at right angles to the first circle.

The last column in the numerical data section, (S), indicates the publication source from which the data in that entry were obtained. Lowercase letters refer to the references listed in the next column, and an asterisk (*) indicates that some of the values have not appeared in previous entries and were calculated by us.

References.—For each item tabulated the name in italics is, to the best of our knowledge, that of the worker who made the measurements. The references corresponding to the letters in column (S) are then keyed by author and year.

Remarks.—Information concerning the age of the rocks is listed here, together with information of value in assessing the reliability of the results for paleomagnetic purposes. The absence of any remarks under the following headings indicates that this particular information was not available.

(1) *Age*—If no specific reference is given for the statements, the age specified was obtained from the principal reference listed in the previous column.

(2) *Sampling*—The remarks under this heading give the areal and stratigraphic extent of the sampling as well as the manner in which statistics were applied to the data.

(3) *Stability*—Any field or laboratory tests that indicate stability or instability are noted here. If no entry is present, no specific test for stability has been reported. Even if a stable component of magnetization is indicated, it does not necessarily follow that the results have paleomagnetic applicability.

(4) *Reversals*—For those studies that show mixed polarities, remarks concerning the number of samples in each group, their stratigraphic relationship, and other relevant observations are noted.

(5) *Other*—Remarks that do not come under the above headings are listed here.

ПОЯСНЕНИЕ К ТАБЛИЦАМ

Приведенные таблицы состоят из трех частей: числовых даных, списка литературы из каторой эти данны почеркнуты, и соответствующих примечаний. Каждое исследованые помечено серийным номером и содержит данные по обрасцу толко из одного источника; порядок последовательности номеров значения не имеет. Подчеркнытые данные представляют материял из каторого были взяты числовые величины, помещенные в тексте следующей части; они дают числовые величины для каждаго столбца, кагда эти величины приведены или могут быть вычислены по данным источника. Ниже дается детальное пояснение что находится под заголовком в каждом столбце.

Rocks sampled.—Название формации или её иное описателное обозначение, вместе с указанием на представленный петрологический тип, даны в этом столбце. Результаты по данному континенту сгруппированы в каждой секции вместе, причем названия континентов набраны жирнбім шрифтом.

No.—В этом стольце дается номер для данных по каждому обрасцу. В каждом разделе таблици номера начинается сначала. Таким образом, полное серийное обозначение цифровых данных по обрасцу состоит из заглавной буквбі, обозначающей раздел талицбі и следующего за буквой номера, на пример **A 3** и **D 4**.

Locality.—В двух колонках под этим обозначением дана географическая жирота и долгота места отбора обрасцов. жирота (*Lat.*) дана в градусах к северу (*N*) или к югу (*S*) от экватора, а долгота (*Long.*) к востоку (*E*) или запасу (*W*) от Гринвича.

Magnetic direction.—В этом столбце дается направление намагниченості и статистиче

ьские данныя. Значение склонения среднего направления намагниченості (*Decl.*) дается в градусах на восток од географического меридиана, а значение наклонения среднего направления намагниченості (*Incl.*) в градусах от горизонтали, со знаком минус для направления вверх от неё. Введена поправка за наклонение пласта. Приведенные статистические данныя указивают "круг сходимости" каторий содержит в себя положение средней точки с вероятности 95% (α_{95}), мера точности (k) и число единичных векторов (*N*), использованых при вычислении этих величины. Величины k и N относятся к направлениям магнитного поля всех тех данных, каторые указывают значения направления намагниченості. В остальных случаях данные k и N относятся к статистическому анализу групи положений полюса.

Pole position.—В этих столбцах даны положения полюсов теоретического геоцентрического диполя (виртуальный магнетный полюс) соответствующий координатам места отбора образцов и среднему значению направления намагниченості. Положения полюсов даны в колонках (*Lat.*) и (*Long.*), так же как это было сделано в стольце для координат места отьора образцов. у каждаго магнитного диполя имеется северный и южныи полюс а данныя в столбце относятся к полюсу находящему в северному полушарю. В столбце (*P*) каторий дает значения поларности полюса, буква *S* обозначает направление намагниченості, соответствующее южному магнитному полюсу в северном полушарии (современое магнитное поле этого типа), а буква *N* указивает, что направление намагниченості соответствует северному магнитному полюсу в северном полушарии. Полюса, вычисленные из данных по образцам имеющие приблизительно противположные или обращенные полярности обозначены буквой (*M*).

(δm) и (δp) обозначают, соотвественно, величину в градусах, болшой и малой полуосей 95-ти процентного овала доверия, относительного среднего положения полюса, соответствующего величине α_{95} относительно среднего магнитного направления по каждому определению. δ_p измеряется по окрушности болжого круга, проходящего через место отборки образца и среднее положение полюса, а δ_m вдоль окружности большого круга перпендикулярного к первому кругу. В последней колонке (*S*) числовыи данных указан источник, из каторого были взятыі числовые данные. Строчные буквы относятся к источникам указанным в следующей колонке, а символ (*) обозначает что некаторые из величин указанные в предидущем столбце не были даны источником, а были вычислены нами.

References.—имя, набраное курсивом относится, согласно наиболее доставерным сведениим, каторые мы могли достать, к работнику, производившему измерения. Затем даны есылки соглпасно буквам в столбце (*S*).

Remarks.—В этом столбце даны сведения, относящиеся к возрасту пород, вместе с оценком достоверности этих сведений для целей палеомагнетизма. Отсутствие примечаний под следующим заглавками указивает на отсутствие источников, из каторых нужные сведения могли быть почеркнуты.

(*1*) *Age.*—Здесь данны замечания о возрасте изучаемых пород. Отсутствие таковых замечаний означает, что данные о возрасте были получены из гловного источника, указаного в предидущем столбце.

(*2*) *Sampling.*—Замечания под этим заголовком дают ареал и стратиграфические указания отобранных образцов, а такше и метод статистической обработки данных.

(*3*) *Stability.*—Здесь указаны полевые или лабораторные иследования магнитной стабилности. Отсутствие данных по иследованию на стабильность указывает, что сведений по этому новоду не имеется. Даже, если указан устоичивый компонент намагниченности, это еще не значит что результаты имеют палеоматнетную значимость.

(*4*) *Reversals.*—Здесь даны замечания относительно образцов в каждой групе, их стратиграфические взаимоотощения и другие, относящиеся к делу наблюдения для случаев смешанной поларности.

(*5*) *Other.*—Заметки и примечания, каторые не входят в предыдущие столбцы.

TABLE 1.—PALEOMAGNETIC DATA

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION A POST EOCENE																
EUROPE																
MT. ETNA LAVAS	1	37½N	15 E	4½	56	2	—	11	—	—	S	—	—	a	*Chevallier* a Hospers, 1955	*Age:* 394 B.C. to A.D. 1911 *Sampling:* 3 to 9 oriented samples were taken from 11 historic lava flows. The last 2 directions agree with observatory data.
	2	**37½N**	**15 E**	**4½**	**56**	**6½**	**50**	**11**	**86 N**	**126 E**	**S**	**9½**	**7**	*		
	3	—	—	—	—	—	—	11	86½N	125½E	S	10	7½	b	b Irving, 1959 c Chevallier, 1925	
CARTHAGE FIRED CLAYS	**4**	**37 N**	**10 E**	**350**	**54½**	**—**	**—**	**—**	**88 N**	**155 W**	**S**	**—**	**—**	*	*Thellier and Thellier* a Thellier and Thellier, 1951	*Age:* 2 dates, 146 B.C. and 300 A.D. *Sampling:* 9 oriented samples from 2 kilns (146 B.C.) and 9 oriented samples from one kiln (300 A.D.) were measured. Values in entry (4) are based on the average declination and inclination given on p. 1478, Ref. a. *Other:* $N = 2$ is not sufficient for calculating Fisher statistics.
PRÄSTMON VARVES	5	63 N	17½E	357½	73½	3½	—	46	—	—	S	—	—	a	*Bancroft* a Hospers, 1955 b Bancroft, 1951	*Age:* 0 to A.D. 1000 *Sampling:* 46 "groups" of specimens were measured.
	6	**63 N**	**17½E**	**357½**	**73½**	**3½**	**42**	**46**	**86 N**	**150 W**	**S**	**6**	**5½**	*		
BRITISH FIRED CLAYS	**7**	**52 N**	**0 E**	**0**	**66½**	**2½**	**242**	**14**	**87 N**	**180 E**	**S**	**4**	**3½**	*	*Cook and Belshé* a Cook and Belshé, 1958	*Age:* First to 15th centuries A.D. *Sampling:* 10 archaeological fired clays were measured from the 1st through 4th centuries and 4 from the 12th, 13th, and 15th centuries. Values for entry (7) are based on data scaled from Fig. 3, Ref. a, which gives declination and inclination values corrected to Cambridge, England. *Other:* Statistical values have no rigorous significance because individual measurements do not represent independent points in time.
ÅNGERMAN RIVER VARVES	**8**	**63 N**	**17½E**	**2**	**74½**	**4½**	**34**	**29**	**88 N**	**150 E**	**S**	**8**	**7½**	*	*Griffiths* a Griffiths, 1955 b Irving, 1959	*Age:* 1100 B.C. to A.D. 750, based on Liden's varve chronology *Sampling:* About 150 samples from two localities a few kilometers apart were measured. These were averaged into 29 groups, each representing about 100 years. Data for the calculations of entry (8) were scaled from Fig. 3(b), Ref. a.
	9	—	—	—	—	—	—	—	88½N	160 E	S	13	12	b		
ICELAND LAVAS	**10**	**—**	**—**	**—**	**—**	**6**	**31**	**21**	**89½N**	**54 E**	**S**	**6**	**6**	*	*Brynjólfsson* a Brynjólfsson, 1957	*Age:* Postglacial (3400 B.C. to 1950 A.D.) *Sampling:* 21 flows were sampled covering about 5000 years. Entry (10) is based on pole-position data for

ICELAND LAVAS	10	—	—	—	—	6	31	21	89½N	54 E	S	6	6	*	*Brynjólfsson* a Brynjólfsson, 1957	*Age:* Postglacial (3400 B.C. to 1950 A.D.) *Sampling:* 21 flows were sampled covering about 5000 years. Entry (10) is based on pole-position data for each flow scaled from Fig. 3, Ref. a. *Stability:* The samples were partially demagnetized in 140 oersted A.C. fields before measurement.
ICELAND LAVAS	11	—	—	1	74	8	—	8	—	—	S	—	—	a	*Hospers* a Hospers, 1955	*Age:* Postglacial *Sampling:* 8 flows were measured covering a period of at least 4000 years. The value of k for entry (12) was calculated by the approximate formula.
	12	64 N	19 W	1	74	8	36	8	86½N	153 E	S	15	13½	*		
SWEDISH VARVES	13	—	—	—	—	12½	16	10	86 N	98 W	S	12½	12½	*	*Granar* a Granar, 1959	*Age:* Glacial and postglacial *Sampling:* 10 varve sections were sampled over a lateral distance of 800 km. Fisher statistics were applied to the pole positions calculated from magnetic direction and locality data on p. 27, Ref. a.
CHAINE DES	14	—	—	353	62	—	—	10	84 N	50 W	S	—	—	a	*Roche*	*Age:* late Pleistocene
PUYS LAVAS	15	45½N	3 E	353	62	—	—	10	84½N	106 W	S	—	—	*	a Roche, 1958	*Sampling:* 10 flows were sampled.
ICELAND LAVAS	16	64½N	22 W	181	−75	7	—	51	—	—	N	—	—	a	*Hospers* a Hospers, 1955	*Age:* early Quaternary *Sampling:* The samples were collected over a lateral extent of 125 km.
	17	64½N	22 W	181	−75	7	9	51	87½N	150 E	N	13	11½	*		
PLATEAUX	18	—	—	206	−63½	—	—	8	72 N	100 E	N	—	—	a	*Roche*	*Age:* early Quaternary
BASALTS	19	45½N	3 E	206	−63½	—	—	8	71½N	84 E	N	—	—	*	a Roche, 1958	*Sampling:* 8 flows were sampled.
FRENCH LAVAS	20	45 N	3½E	197	−62½	13	—	6	—	—	N	—	—	b	*Roche* a Roche, 1951 b Hospers, 1955	*Age:* Pliocene and Pleistocene. The oldest flows are Villafranchien. *Sampling:* 6 flows were sampled.
	21	45 N	3½E	197	−62½	13	28	6	78½N	93 E	N	20	15½	*		
ICELAND LAVAS	22	—	—	—	—	—	19	33	77½N	74 E	S	—	—	a	*Sigurgeirsson* a Sigurgeirsson, 1957	*Age:* Pliocene and Pleistocene *Sampling:* 33 lava flows from three normally magnetized groups and 26 flows from three reversely magnetized groups were sampled. Fisher statistics were applied to the pole positions calculated from each flow. The values of α_{95} for entries (23) and (25) were calculated by the approximate formula. *Stability:* A.C. demagnetization of 110 oersted reduced the scatter in the determinations, especially in the reversed groups.
	23	—	—	—	—	5½	19	33	77½N	74 E	S	5½	5½	*		
	24	—	—	—	—	—	15	26	88 N	149 E	N	—	—	a		
	25	—	—	—	—	7	15	26	88 N	149 E	N	7	7	*		

were scaled from Fig. 3(b), Ref. a.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION A—*Continued*																
CHELEKAN SEDIMENTS	26	—	—	12	37	—	—	—	—	—	S	—	—	a	*Khramov*	*Age:* Pliocene and Pleistocene
	27	**39 N**	**53 E**	**12**	**37**	**—**	**—**	**—**	**69 N**	**161 W**	**S**	**—**	**—**	*	a Khramov, 1957	*Sampling:* 650 oriented samples were measured from localities as far as 170 km from each other.
	28	—	—	196	−30	—	—	—	—	—	N	—	—	a		*Other:* In 4 bore holes 2–4 km apart, transition from north-seeking to south-seeking magnetizations occurs at the same level. Samples from several "normal" and "reversed" zones are included in the average directions cited. Additional details appear in Khramov, 1958.
	29	**39 N**	**53 E**	**196**	**−30**	**—**	**—**	**—**	**63 N**	**163 W**	**N**	**—**	**—**	*		
FRENCH LAVAS	30	45½N	3 E	176½	−51	14	—	5	—	—	N	—	—	b	*Roche*	*Age:* Miocene and Pliocene. The oldest flows are Pontian.
	31	**45½N**	**3 E**	**176½**	**−51**	**14**	**27**	**5**	**76 N**	**164 W**	**N**	**19**	**12½**	*	a Roche, 1951	
	32	—	—	—	—	—	—	5	73 N	167½W	N	19	13½	c	b Hospers, 1955	*Sampling:* 5 flows were sampled.
															c Irving, 1959	
ICELAND LAVAS	33	65 N	20 W	1½	78	5½	—	102	—	—	M	—	—	a	*Hospers*	*Age:* Miocene
	34	**65 N**	**20 W**	**1½**	**78**	**5½**	**7**	**102**	**88 N**	**10 W**	**M**	**10½**	**10**	*	a Hospers, 1955	*Sampling:* An average of 25 flows from each of 4 normal and reversed zones were measured.
VOGELSBERG BASALTS															*Angenheister*	*Age:* Cited as Miocene
Normal flows	35	—	—	8	57	8	—	29	—	—	S	—	—	a	a Angenheister, 1956	*Sampling:* More than 200 oriented samples were collected from 42 flows. The values of k for entries (36), (38), and (40) were calculated by the approximate formula.
	36	50½N	9½E	8	57	8	11	29	76 N	163½	S	11½	8	*		
Reversed flows	37	—	—	188½	−60	15	—	13	—	—	N	—	—	a		
	38	50½N	9½E	188½	−60	15	7	13	79 N	155 E	N	22	17	*		
All flows	39	—	—	8½	57½	6	—	42	—	—	M	—	—	a		
	40	**50½N**	**9½E**	**8½**	**57½**	**6**	**13**	**42**	**76½N**	**160 E**	**M**	**8½**	**6½**	*		
ENGLISH NORTH WEST DYKES	**41**	**55½N**	**3 W**	**179½**	**−73**	**15½**	**16**	**7**	**87 N**	**8 W**	**N**	**28**	**25**	*	*Bruckshaw and Robertson*	*Age:* Cited as Oligocene or Miocene in Refs. a and b
	42	55½N	3 W	174½	−73½	15½	—	7	—	—	N	—	—	b		*Sampling:* The samples were collected from 4 dikes at 7 sites. The sites covered an area of about 50 by 140 miles. Values for entry (41) were based on data scaled from Fig. 7B, Ref. a.
	43	—	—	—	—	—	—	7	86 N	76 W	N	25	21	c	a Bruckshaw and Robertson, 1949	
															b Hospers, 1955	
															c Irving, 1959	
LIMAGNE BASALT	44	—	—	180	−73	—	—	—	—	—	N	—	—	a	*Roche*	*Age:* Cited as Aquitanian (lower Miocene)
	45	—	—	—	—	—	—	—	77 N	3 E	N	—	—	b	a Roche, 1950b	*Sampling:* One locality was sampled.
	46	**46 N**	**3 E**	**180**	**−73**	**—**	**—**	**—**	**77 N**	**3 E**	**N**	**—**	**—**	*	b Irving, 1959	
FRENCH INTRUSIVE ROCKS	47	46 N	3 E	201	−57	11	—	9	—	—	N	—	—	a	*Roche*	*Age:* Cited as Oligocene in Ref. a
	48	**46 N**	**3 E**	**201**	**−57**	**11**	**24**	**9**	**72 N**	**114 E**	**N**	**15½**	**11½**	*	a Hospers, 1955	
	49	—	—	—	—	—	—	—	73 N	119 E	N	16	12	c	b Roche, 1950a	
															c Irving, 1959	
NORTH AMERICA																
NEW ENGLAND	50	43 N	72½W	355½	51½	10	—	6	—	—	S	—	—	a	*Johnson, Murphy,*	*Age:* 13,000 to 7000 B.C.

SIVE ROCKS	**48** 49	**46 N** —	**3 E** —	**201** —	**−57** —	**11** —	**24** —	**9** —	**72 N** 73 N	**114 E** 119 E	**N** N	**15½** 16	**11½** 12	* c	a Hospers, 1955 b Roche, 1950a c Irving, 1959	
NORTH AMERICA																
NEW ENGLAND VARVES	50	43 N	72½W	355½	51½	10	—	6	—	—	S	—	—	a	*Johnson, Murphy, and Torreson*	*Age:* 13,000 to 7000 B.C. *Sampling:* The measurements were averaged into 6 groups, each representing 1000 years, and the calculations were based on the average directions for these 6 groups.
	51	**43 *N***	**72½W**	**355½**	**51½**	**10**	**50**	**6**	**79 N**	**127 E**	**S**	**13½**	**9**	*	a Hospers, 1955	
	52	—	—	—	—	—	—	—	80½	132 E	S	13½	9½	b	b Irving, 1959	
															c Johnson, Murphy, and Torreson, 1948	
CANADIAN BASALTS	53	—	—	348½	75	4	28	46	—	—	M	—	—	a	*Du Bois*	*Age:* Cited as late Tertiary in Ref. a
	54	**61 N**	**134½W**	**348½**	**75**	**4**	**28**	**46**	**85 N**	**145 E**	**M**	**6½**	**7½**	*	a Du Bois, 1959a	*Sampling:* 29 samples were collected at 60½ N, 135 W; 9 at 60 N, 130½ W; 6 at 63 N, 138 W; and 4 at 59½ N, 134 W; 2 widely divergent samples from the second group were discarded. There is no indication of the number of flows sampled. Location data for entry (54) are averages of the above 4 localities. *Reversals:* All the samples from the first and third groups were reversed with respect to the present field direction.
NEROLY FORMATION (sediments)	55	—	—	7	58	3	—	29	—	—	S	—	—	a	*Doell*	*Age:* Magnetization is parallel before correcting for post early Pleistocene folding with dips of 50°; therefore the magnetization was acquired post early Pleistocene, although the rocks were deposited in late Miocene time. *Sampling:* 29 oriented samples were collected from 3 areas about 50 miles apart. The value of *k* for entry (56) was calculated using the approximate formula. *Stability:* Partial heat demagnetization to 100°C. decreased the scatter in directions.
	56	**37½N**	**122 W**	**7**	**58**	**3**	**29**	**29**	**85 N**	**42 W**	**S**	**4**	**3**	*	a Doell, 1955a b Doell, 1956	
PAYETTE FORMATION (sediments)	**57**	**43 N**	**115½W**	**1½**	**62½**	**2½**	**258**	**13**	**88½N**	**60 W**	**S**	**4**	**3**	*	*Torreson et al.* a Torreson, Murphy, and Graham, 1949	*Age:* Cited as Pliocene in Ref. a although now regarded by U. S. Geological Survey as Miocene and Pliocene(?) *Sampling:* 13 oriented samples were collected at one site. Entry (57) was based on data given in Table 4, Ref. a.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION A—*Continued*																
COLUMBIA RIVER BASALTS															*Campbell and Runcorn* a Campbell and Runcorn, 1956 b Irving, 1959	*Age:* Miocene *Sampling:* 73 separate flows are included in the calculations out of a total of 114 examined in the area. 7 localities were sampled over an area of 300 by 200 miles. Values for entry (62) were obtained by averaging the values for declination, inclination, and k given in entries (58) and (60). The value for α_{95} was then calculated by the approximate formula.
Normal flows	58	—	—	11½	73½	7½	5	44	—	—	S	—	—	a		
	59	46½N	120 W	11½	73½	7½	5	44	75½N	97 W	S	13½	12½	*		
Reversed flows	60	—	—	177	−66	10½	4	29	—	—	N	—	—	a		
	61	46½N	120 W	177	−66	10½	4	29	87 N	170 W	N	17½	14½	*		
All flows	**62**	**46½N**	**120 W**	**3½**	**69½**	**8**	**4**	**73**	**83 N**	**105 W**	**M**	**13½**	**11½**	*		
	63	—	—	—	—	—	—	—	87 N	40 E	M	4½	3½	b		
ELLENSBURG FORMATION (sediments)	64	46½N	120½W	4	66½	10½	11	19	86½N	73 W	S	17½	14	*	*Torreson et al.* a Torreson, Murphy, and Graham, 1949 b Graham, 1949	*Age:* Cited as Miocene in Ref. a. Now regarded as late Miocene and early Pliocene by U. S. Geological Survey. *Sampling:* 23 oriented samples were collected at one site. The calculations for entry (64) were based on data scaled from Fig. 11, Ref. b, and those for entry (65) were based on data given in Table 4, Ref. a. *Stability:* Stability is indicated by the application of Graham's conglomerate test. *Reversals:* One sample showed reversed polarity.
	65	**46½N**	**120½W**	½	**68½**	**9**	**12**	**23**	**85 N**	**115 W**	**S**	**15½**	**13**	*		
ARIKAREE FORMATION (sediments)	**66**	**44 N**	**103 W**	**66**	**69**	**25½**	**3**	**21**	**47 N**	**49 W**	**S**	**43½**	**37**	*	*Torreson et al.* a Torreson, Murphy, and Graham, 1949	*Age:* Cited as Miocene. *Sampling:* 21 oriented samples were collected at one site. Entry (66) was based on data given in Table 4, Ref. a. *Other:* The scatter in directions is extreme.
DUCHESNE RIVER FORMATION (sediments)	**67**	**40½ N**	**110 W**	**2**	**65**	**5**	**14**	**85**	**83 N**	**99 W**	**S**	**8**	**6**	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Tertiary. *Sampling:* 85 specimen measurements were made on 24 oriented samples.
SOUTH AMERICA																
NEUQUEN LAVAS															*Creer*	*Age:* Cited as Quaternary.

SOUTH AMERICA															
NEUQUEN LAVAS														*Creer*	*Age:* Cited as Quaternary.
Normal flows	−68	—	—	350	−46	8	31	12	—	—	S	—	—	a Creer, 1958	*Sampling:* In all these entries, N is the number of specimen measurements which were made on 20 oriented samples collected from 10 flows (both normal and reversed). Lateral sampling extent was about 200 by 400 km. Entry (74) was based on data scaled from plots in Ref. a. Entry (75) is based on the "secular variation precision" $k' = 30$ and the number of flows sampled, and α_{95} was calculated by the approximate formula.
	69	38 S	70 W	350	−46	8	31	12	76½N	112 W	S	10	6	b Irving, 1959	
Reversed flows	−70	—	—	188	64	5	17	46	—	—	N	—	—	a	
	71	38 S	70 W	188	64	5	17	46	80½N	76 E	N	8½	6½	*	
	72	—	—	1	−61	5	15	58	—	—	M	—	—	a	
All flows	73	—	—	—	—	—	—	—	83 N	126 E	M	7	6	b	
	74	38 S	70 W	3	−69	3½	15	65	75 N	102 E	M	6	5	*	
	75	**38 S**	**70 W**	**3**	**−69**	**8**	k′	**10**	**75 N**	**102 E**	**M**	**13½**	**11½**	*	*Stability:* The scatter in directions was very greatly reduced by partial A.C. demagnetization at 250 oersted.
AUSTRALIA															
NEWER VOLCANICS OF VICTORIA	76	38 S	143½E	8	−59½	8	35	13	—	—	S	—	—	a	*Irving and Green*	*Age:* Pliocene to Recent
	77	38 S	143½E	177	60	6½	31	16	—	—	N	—	—	a	a Irving and Green, 1957	*Sampling:* At least 2 oriented samples were collected at each of 32 sites covering an area of 50 by 150 miles. The values cited here for k are between-site precision parameters (see Watson and Irving, 1957, p. 296–297).
	78	**38 S**	**143½E**	**3½**	**−60**	**5**	**37**	**32**	**86½N**	**78 W**	**M**	**7**	**5½**	a	
NEW ZEALAND IGNIMBRITES	79	—	—	—	—	—	—	—	79 N	35 E	S	—	—	b	*Hatherton*	*Age:* Cited as Pliocene
	80	**37½S**	**175 E**	**351**	**−65**	—	—	—	**79 N**	**26 E**	**S**	—	—	*	a Hatherton, 1954 b Irving, 1959	*Sampling:* 52 oriented samples were measured as well as 65 samples cut from vertical drill cores. Values for entry (80) were based on the inclination and declination data given in Table 3, Ref. a. Lateral sampling extent was 10 miles.
ASIA															
JAPANESE FIRED CLAYS	**81**	**35½N**	**140 E**	**358**	**52**	**3**	**58**	**45**	**86½**	**5 W**	**S**	**4**	**3**	*	*Watanabe* a Watanabe, 1958	*Age:* From 5600 to 4400 B.C. and 300 to 1800 A.D. *Sampling:* Data from 43 archaeological fired clays and 2 dated lava flows were reported, 22 in the period 300 to 1800 A.D. and 23 in the period 5600 to 4400 B.C. Values for entry (81) were based on data scaled from Figs. 1 and 2, and cited in Table 1, Ref. a.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION A—*Continued*																
NORTH IZU AND HAKONE VOLCANIC ROCKS	82	—	—	—	—	7	11	42	78 N	45½E	M	7	7	°	*Nagata et al.*	*Age:* Quaternary
	83	—	—	—	—	—	—	—	76½N	37½E	M	10½	7	b	a Nagata, Akimoto, Uyeda, Shimizu, Ozima, Kobayashi, and Kuno, 1957 b Irving, 1959	*Sampling:* 42 flows were measured more or less uniformly throughout the Quaternary. The calculations for entry (82) are in the form of a Fisher analysis on the pole positions for each group of flows cited in Table 2, Ref. a. Since values of α_{95} for the 42 sets of data vary widely, the overall value of α_{95} listed has no rigorous statistical significance. *Stability:* Stability was established by extensive and elaborate laboratory tests. *Reversals:* 9 flows near the bottom of the Quaternary show reversed polarity.
JAPAN VOLCANIC ROCKS	84	36 N	138 E	11	61½	11	5	39	79½N	173 W	M	17	13	°	*Matuyama* a Matuyama, 1929	*Age:* The rocks sampled are Tertiary and younger. *Sampling:* 39 oriented samples were collected at 35 localities in Honsyû, Kyûsyû, Tyôsen, and Manchuria, the majority being collected in Honsyû. The data used in the calculations for entry (84) were scaled from a figure on p. 204, Ref. a. The latitude & longitude of the sampling area are an average for Honsyû; thus there may be somewhat more dispersion in the values given than there would have been at a single sampling area. *Reversals:* About half the samples measured showed reversed polarity.
JAPAN VOLCANIC ROCKS	85	36 N	138 E	359	47½	11	18	11	82½N	35 W	S	14½	9½	°	*Kumagai et al.* a Kumagai, Kawai, and Nagata, 1950	*Age:* Pleistocene to Recent *Sampling:* 11 lava flows were sampled. Data for the calculation of entry (85) were taken from Tables I and II, Ref. a (nos. 1, 2, 3, 5, 6, 7, 17, 18, 19, 20, 21). The Narita bed, no. 4, is stated by Kawai (1954, p. 209) to be unstable and was not included here. Coordinates for the sampling area are averages.
JAPAN VOLCANIC ROCKS	86	36 N	138 E	349½	48	12½	37	7	79 N	13 E	M	16	10½	°	*Kumagai et al.* a Kumagai, Kawai	*Age:* Late Tertiary *Sampling:* 7 lava flows were sampled. Data for the

JAPAN VOLCANIC ROCKS	86	36 N	138 E	349½	48	12½	37	7	79 N	13 E	M	16	10½	*	*Kumagai et al.* a Kumagai, Kawai, and Nagata, 1950	*Age:* Late Tertiary *Sampling:* 7 lava flows were sampled. Data for the calculation of entry (86) were taken from Table I, Ref. a. Nos. 9 through 15 were used. No. 8 was discarded because of instability (Kawai, 1954, p. 209). Co-ordinates for the sampling area are averages. *Reversals:* One of these flows showed reversed polarity.
KAWAJIRI BASALT	87	—	—	—	—	—	—	—	80 N	49 E	N	—	—	a	*Asami* a Irving, 1959 b Asami, 1954a c Asami, 1954b	*Age:* early Pleistocene *Sampling:* One flow was sampled. *Other:* Normal and intermediate directions of magnetization were found in closely associated lavas (Ref. c, p. 151).
JAPAN VOLCANIC ROCKS															*Nagata et al.* a Nagata, Akimoto, Shimizu, Kobayashi, and Kuno, 1959	*Age:* late and early Pliocene *Sampling:* 2 basalt lavas were sampled from the upper Pliocene and 2 andesite lavas from the lower Pliocene. Values for calculation of entry (90) are based on direction data given in Table II, Ref. a, and the localities cited in Table 1, Ref. a. Fisher statistics were applied to the 4 pole positions calculated from these data.
upper Pliocene	88	—	—	3	42	2	—	12	79 N	57 W	S	3	2	a		
lower Pliocene	89	—	—	14	52	4	—	10	77 N	109 W	S	6	4	a		
combined	90	—	—	—	—	10	82	4	79½N	88½W	S	10	10	*		
JAPAN VOLCANIC ROCKS															*Nagata et al.* a Nagata, Akimoto, Shimizu, Kobayashi, and Kuno, 1959	*Age:* late, middle, and early Miocene *Sampling:* A dolerite sheet, andesite sheet, and 2 andesite lavas were sampled from the upper and middle Miocene and a dolerite sheet and 2 andesite lavas from the lower Miocene. Values for calculation of entry (93) are based on direction data given in Table II, Ref. a, and the localities cited in Table I, Ref. a. Fisher statistics were applied to the 7 pole positions calculated from these data. *Reversals:* 2 units from the upper and middle Miocene and one from the lower Miocene had reversed polarity.
upper and middle Miocene	91	—	—	27	59	3	—	32	68 N	152 W	M	5	4	a		
lower Miocene	92	—	—	32	40	11	—	20	59 N	113 W	M	13	8	a		
combined	93	—	—	—	—	22	8	7	73 N	144 W	M	22	22	*		

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION B PRECAMBRIAN																
EUROPE TORRIDONIAN SANDSTONE															*Irving and Runcorn* a Creer, Irving, and Runcorn, 1954 b Runcorn, 1955b c Runcorn, 1955a d Day and Runcorn, 1955 e Irving and Runcorn, 1957 f Creer, Irving, and Runcorn, 1957 g Irving, 1957a	*Age:* late Precambrian, overlying Lewis gneiss and overlain by Lower Cambrian rocks. Entries (1) through (12) are Aultbea, Applecross, and top Diabaig formations. Entries (13) through (16) are lower Diabaig. Aultbea overlies Applecross which overlies Diabaig. *Sampling:* Lateral extent, 60 miles; vertical extent, 10,500 feet in upper Torridonian and 1,900 feet in lower Torridonian. N refers to the number of sampling sites; the mean direction at each site was used in the statistical analysis. 11 sites in the upper Torridonian were magnetized in directions oblique to the mean direction of the remaining 81 sites; these, as well as 6 sites with directions parallel to the present field, were not used in the statistical analysis. "Normal" and "reversed" groups of sites differ significantly in their mean directions but were combined in entry (12) for mean axial direction of magnetization. *Other:* The change in magnetic direction between upper and lower Torridonian occurs stratigraphically in less than 300 feet with no evidence of unconformity. *Stability:* Torridonian pebbles in conglomerates of later age have randomly oriented directions of magnetization. Directions of magnetization are scattered in beds showing penecontemporaneous deformation (Ref. e, p. 93). Directions of magnetization of samples collected from a fold of Caledonian age are parallel only after correcting for dip, indicating stability at least since the Caledonian (Ref. e, p. 92–93). *Reversals:* About 16 zones in the upper Torridonian have magnetizations which alternately lie in approximately opposite directions. In two cases obliquely magnetized zones occur stratigraphically between two reversed zones. No reversals occur in the lower Torridonian.
Upper	1	—	—	295	−34	—	—	—	—	—	—	—	—	a		
	2	58 N	6 W	295	−34	—	—	—	3 N	53 E	N	—	—	*		
	3	58 N	6 W	294	−28	9	9	28	—	—	—	—	—	e		
	4	58 N	6 W	294	−28	9	9	28	1 N	56 E	N	10	5½	*		
	5	—	—	127	52	—	—	—	—	—	—	—	—	a		
	6	58 N	6 W	127	52	—	—	—	11½N	37 E	S	—	—	*		
	7	58 N	6 W	129	51	5	15	53	—	—	—	—	—	e		
	8	58 N	6 W	129	51	5	15	53	10 N	36 E	S	6½	4½	*		
	9	—	—	304	−38	—	—	—	—	—	—	—	—	a		
	10	58 N	6 W	304	−38	—	—	—	2 N	45 E	M	—	—	*		
	11	—	—	—	—	—	—	—	5 N	49 E	—	8	5	b		
	12	**—**	**—**	**123**	**44**	**5**	**12**	**61**	**6 N**	**43 E**	**M**	**6**	**4**	**f**		
Lower	13	—	—	—	—	—	—	—	37 N	109 W	—	—	—	d		
	14	—	—	307	34	7	40	—	—	—	—	—	—	c		
	15	—	—	307	34	7	40	13	35 N	112 W	S	8	5	f		
	16	**58 N**	**6 W**	**307**	**34**	**7**	**—**	**—**	**35 N**	**118 W**	**S**	**8**	**5**	**g**		
LONGMYNDIAN (Sediments)	17	—	—	111	19	—	—	—	—	—	—	—	—	a	*Creer*	*Age:* late Precambrian, possibly equivalent to Tor-

LONGMYNDIAN (Sediments)	17	—	—	111	19	—	—	—	—	—	—	—	—	a	*Creer*	*Age:* late Precambrian, possibly equivalent to Torridonian
	18	53 N	3 W	111	19	—	—	—	4½N	115½W	M	—	—	*	a Creer, Irving, and Runcorn, 1954	*Sampling:* Lateral extent, 20 miles. Vertical extent, probably several thousand feet. 40 samples were collected at 12 sites. Statistical analysis apparently is of directions of individual samples, not of mean site directions (Ref. d).
	19	—	—	—	—	—	—	—	2 N	118 W	—	13	7	b	b Runcorn, 1955b	
	20	—	—	114	29	12	5	40	2 N	120 W	M	13	7	c	c Creer, Irving, and Runcorn, 1957	
	21	**53 N**	**3 W**	**114**	**29**	**12**	**5**	**40**	**1½N**	**121 W**	**M**	**13**	**7**	*	d Creer, 1957b	*Stability:* Magnetization unchanged in 300 oersted alternating field and also unchanged after 1 year's random orientation in the laboratory.
																Reversals: Present, but number of alternating zones not known.
NORTH AMERICA																
CHEQUAMEGON SANDSTONE	22	—	—	30	74	6	—	15	69 N	47 W	—	—	—	a	*Du Bois*	*Age:* late Keweenawan
	23	**47 N**	**88½W**	**30**	**74**	**6**	**36**	**15**	**68 N**	**47 W**	**S**	**11**	**10**	*	a Du Bois, 1957	*Sampling:* *N* is listed as number of "specimens." The value of *k* for entry (23) was calculated by the approximate formula.
JACOBSVILLE SANDSTONE	24	—	—	250	−11	13	—	15	14 N	10 E	N	—	—	a	*Du Bois*	*Age:* late Keweenawan, older than Chequamegon
	25	**47 N**	**88½W**	**250**	**−11**	**13**	**8**	**15**	**17½N**	**13 E**	**N**	**13**	**6½**	*	a Du Bois, 1957	*Sampling:* *N* is listed as number of "specimens." The value of *k* for entry (25) was calculated by the approximate formula.
FREDA SANDSTONE AND NONESUCH SHALE	26	—	—	—	—	—	—	—	20 N	165 E	—	—	—	a	*Du Bois*	*Age:* late Keweenawan, older than Jacobsville; conformable on Copper Harbor conglomerate
	27	—	—	285	−1	3	—	68	9 N	169 E	—	—	—	b	a Du Bois, 1955	*Sampling:* Lateral extent, probably about 100 miles. *N* is number of "specimens." Value of *k* for entry (28) was calculated by the approximate formula.
	28	**46½N**	**88½W**	**285**	**−1**	**3**	**32**	**68**	**10 N**	**170 E**	**S**	**3**	**1½**	*	b Du Bois, 1957	*Stability:* Fold tests indicate stability.
COPPER HARBOR (Sediments and Lava flows)	29	—	—	294	32	7	—	25	30 N	176 E	—	—	—	a	*Du Bois*	*Age:* late Keweenawan
	30	**47 N**	**88½W**	**294**	**32**	**7**	**16**	**25**	**29 N**	**176 E**	**S**	**8**	**4½**	*	a Du Bois, 1957	*Sampling:* 13 samples are from lava flows, and 12 are from sediments. The value of *k* for entry (30) was calculated by the approximate formula.
															b Du Bois, 1955	*Stability:* Copper Harbor(?) basalt and andesite-pebbles in Copper Harbor conglomerates are randomly magnetized. Thermal demagnetization curves (of basalts?) are similar to those for TRM.
																Other: Sediments have 10° smaller inclination than lavas above and below them (Ref. b, p. 507).

two reversed zones. No reversals occur in the lower Torridonian.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION B—*Continued*																
PORTAGE LAKE LAVA SERIES	31	—	—	282	41	4	—	31	25 N	170 W	—	—	—	a	*Du Bois*	*Age:* middle Keweenawan; conformably overlain by Copper Harbor
	32	47 N	88½W	282	41	4	40	31	25 N	170 W	S	5	3	*	a Du Bois, 1957	*Sampling:* N is number of specimens. The value of k for entry (32) was calculated by the approximate formula.
																Stability: Thermal decay curves are similar to those for TRM.
MICHIGAN DIABASE DIKES	33	46½N	88½W	82	−86	1	82	36	45 N	99 W	N	2	2	a	*Graham*	*Age:* late Precambrian. Overlain in adjacent localities by Jacobsville and by flat-lying Cambrian sediments. Probable age about 1100 million years (James, 1958, p. 40).
															a Creer, Irving, and Runcorn, 1957	*Sampling:* Statistical analysis is based on 36 samples from 2 dikes 8 miles apart. 20 samples from 1.8-foot thick dike span a distance of 3½ feet. 16 samples from 40-foot thick dike span a distance of 15 feet. Samples are from both chilled borders and coarse interiors. Several samples from a third dike agree with these directions; samples from two additional dikes are widely scattered in direction.
															b Graham, 1953	*Stability:* One sample retained direction and 70% of intensity of magnetization in alternating magnetic field of 493 oersted.
																Reversals: Graham (1953, p. 252–254) believed dikes may have undergone self-reversal in field essentially parallel to present field, but laboratory evidence for self-reversal is not conclusive.
ADIRONDACK METAMORPHIC ROCKS	34	—	—	—	—	—	—	—	44 N	74 W	N	—	—	a	*Balsley and Buddington*	*Age:* Du Bois (Ref. a) believes remanent magnetization was acquired during metamorphism of Grenville age

Adirondack Metamorphic Rocks	34	—	—	—	—	—	—	—	44 N	74 W	N	—	—	a	*Balsley and Buddington* a Du Bois, 1958 b Du Bois, 1957 c Balsley and Buddington, 1954	*Age:* Du Bois (Ref. a) believes remanent magnetization was acquired during metamorphism of Grenville age (ca. 1000 million years). *Other:* Pole entry (34) was calculated on the basis of the report by Balsley and Buddington (1954) that rocks containing titano-hematite as the principal magnetic constituent tend to be vertically and reversely polarized. Co-ordinates were scaled from Fig. 1, Ref. a. Balsley and Buddington (1958, p. 790–792) question the interpretation that the remanent magnetization of these rocks is simple TRM parallel to the field in which the rocks cooled.
Gabbro Intrusive Rocks, Bancroft Area, Ontario															*Hood* a Hood, 1958	*Age:* Precambrian (Grenville province). Radio-isotope ages of igneous rocks in this part of the Grenville province generally range between 1000 and 1200 million years (Shillibeer and Cumming, 1956, p. 56–57; Wilson, 1958, p. 762).
Boulter Intrusive	**35**	**45 N**	**77½W**	**297½**	**55**	**5**	**19**	**43**	**42½N**	**157 W**	**S**	**7**	**5**	**a**		*Sampling:* Statistical analysis is based on directions of 43 specimens from 8 oriented samples. *Stability:* Alternating field decreased scatter in directions of magnetization. *Other:* Gabbro is locally gneissic, approaching metagabbro. Laboratory observations suggest susceptibility anisotropy.
Umfraville Instrusive	36	45 N	78 W	115	42½	7	8	58	—	—	—	—	—	a		*Sampling:* Statistical analysis is based on 58 specimens from 12 oriented samples; anomalously magnetized specimens were not included in the analysis. *Other:* Umfraville intrusive is less metamorphosed than Boulter.
	37	45 N	78 W	115	42½	7	8	58	1 N	22 W	S	9	5½	*		
Thanet Intrusive	38	45 N	77½W	92½	62½	14½	5	28	—	—	—	—	—	a		*Sampling:* Statistics are based on 28 specimens from 6 oriented samples. Locality is 5 miles from Umfraville site. *Stability:* Alternating magnetic field demagnetization decreased scatter in directions of magnetization.
	39	45 N	77½W	92½	62½	14½	5	28	28 N	23½W	S	22	18	*		
Thanet and Umfraville	40	45 N	77½W	110	49	7	6	86	34 N	22 W	S	9	6	a		*Sampling:* Combination of two previous groups.
	41	**45 N**	**77½W**	**110**	**49**	**7**	**6**	**86**	**8½N**	**22 W**	**S**	**9**	**6**	*		
Tudor Intrusive	42	44½N	77½W	327½	11½	9	5	68	—	—	—	—	—	a		*Sampling:* Statistics are based on 68 specimens from 11 oriented samples. Tudor locality is 8 miles south of Thanet.
	43	**44½N**	**77½W**	**327½**	**11½**	**9**	**5**	**68**	**41½N**	**149 E**	**S**	**9**	**4½**	*		

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION B—*Continued*																
SUDBURY INTRUSIVE															*Hood* a Hood, 1958	*Age:* Precambrian. Radio-isotope age of between 1200 and 1800 million years is generally assigned (Russell and others, 1954, p. 307–308; Wetherill and others, 1957, p. 412). Four localities (Azilda, Blezard, Garson, Creighton) spanning 16 miles were sampled on the south side of the Sudbury Basin. One locality (Levack) was sampled on the north side.
Levack	44	46½N	81½W	320½	70	5	20	41	64 N	140½W	S	9	8	a		*Sampling:* Statistics based on 41 specimens from 14 oriented samples; some anomalous specimens were not included.
Azilda	45	46½N	81 W	194½	66½	2½	67	45	—	—	—	—	—	a		*Sampling:* Statistical analysis is based on 45 specimens from 13 oriented samples. Lateral sampling extent is 1.6 miles.
Blezard	46	46½N	81 W	184	71½	1½	86	78	—	—	—	—	—	a		*Sampling:* Lateral sampling extent is 1.4 miles. Statistical analysis based on 78 specimens from 12 oriented samples. *Stability:* Partial demagnetization in 124 oersted A.C. field decreased scatter.
Garson	47	46½N	81 W	155½	59	10	44	6	—	—	—	—	—	a		*Sampling:* Statistical analysis is based on 6 specimens from 2 oriented samples.
Creighton	48	46½N	81 W	171½	57	4	47	26	—	—	—	—	—	a		*Sampling:* Statistics are based on 26 specimens from 6 oriented samples. *Stability:* 125 oersted alternating field had little effect on intensity or direction of magnetization.
South Range of Sudbury, combined	49	46½N	81 W	183	68	1½	49	155	7½N	94½W	S	3	2	a		*Sampling:* Entry (49) is based on all 155 specimens from 33 samples collected at all 4 sites. Entry (51) is based on the mean direction of magnetization at each of 4 sites.
	50	46½N	81 W	183	68	1½	49	155	7½N	82½W	S	3	2	*		
	51	46½N	81 W	174	64	11½	65	4	—	—	—	—	—	*		
Both sides of Sudbury Basin, combined	52	46½N	81 W	245½	82½	—	—	196	38½N	99½W	S	—	—	a		*Other:* The magnetic direction in entry (52) was chosen midway between the direction for Levack on the north rim of the basin and the average direction entry (49) for the sites on the south rim. Entry (53) is based on the following geologic considerations. Thomson (1956, p. 44–45) concludes that the gently dipping north limb and steeply dipping south limb of the Sudbury syncline were largely developed
	53	**46½N**	**81 W**	**300**	**78**	**11**	**—**	**—**	**53 N**	**115 W**	**S**	**21**	**20**	*		

before intrusion; most of the post-intrusion folding was in the south limb. The magnetic evidence sup-

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION B—*Continued*																
SHINUMO QUARTZITE, GRAND CANYON	66	36 N	112 W	288	65	—	—	61	37 N	166 W	S	—	—	a	*Collison and Runcorn* a Collinson and Runcorn, 1960	*Age:* Precambrian; overlies Hakatai shale *Sampling:* Entry (66) is based on 61 specimens from 14 oriented samples. Directions of magnetization show streaking toward direction of present field; in entry (67) a mean direction was estimated excluding directions tending toward the present field direction.
	67	36 N	112 W	246	33	—	—	—	7 N	7 E	N	—	—	a		
HAZEL FORMATION (sediments)	68	—	—	—	—	—	—	—	49 N	175 W	S	—	—	a	*Howell, Martinez, and Statham* a Howell, Martinez, and Statham, 1958	*Age:* Precambrian *Sampling:* Entries (68, 69) are based on 15 samples from flat-lying beds exposed at 5 localities scattered over an area of 2 square miles. Entries (70, 71) based on 37 oriented samples from dipping beds from 9 localities scattered over 20 miles, correction for dip having been made. Entries (69) and (71) are based on data scaled from Fig. 3 and Fig. 4, respectively, in Ref. a. *Stability:* Directions of magnetization of some samples changed 2°–16° in several months. *Other:* Rocks may be slightly metamorphosed.
	69	31 N	105 W	316	56½	6½	35	15	53 N	173 W	S	9½	7	*		
	70	—	—	—	—	—	—	—	60 N	151 E	S	—	—	a		
	71	31 N	105 W	328	37½	17	3	37	59½N	154 E	S	20	12	*		
BELT SERIES McNamara Formation (sediments)	72	47 N	114 W	26	−43	4	30	53	14 N	42 E	S	5	3	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Precambrian *Sampling:* Statistics are based on 53 specimens from 20 oriented samples.
Miller Peak Formation (sediments)	73	47 N	114 W	234	30	7	20	23	11 N	14 E	N	8	4	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Precambrian *Sampling:* Statistics are based on 23 specimens from 14 oriented samples.
Spokane Shale	75	47 N	112 W	232	55	4	18	71	5 N	152 W	S	6	4	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Precambrian; younger than Grinell formation *Sampling:* Statistics are based on 71 specimens from 39 oriented samples.
	76	49 N	114 W	206	39	8	10	19	16 N	41 E	N	10	6	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Precambrian; younger than Grinell formation *Sampling:* Statistics are based on 19 specimens from 5 oriented samples.
Appekunny Argillite	77	49 N	113½ W	223	29	6	15	38	15 N	24 E	N	7	4	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Precambrian; younger than Appekunny argillite *Sampling:* Statistics are based on 38 specimens from 15 oriented samples.

															a Collinson and Runcorn, 1960	5 oriented samples.
Appekunny Argillite	77	49 N	113½ W	223	29	6	15	38	15 N	24 E	N	7	4	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Precambrian; younger than Appekunny argillite *Sampling:* Statistics are based on 38 specimens from 15 oriented samples.
Grinell Formation (sediments)	78	49 N	113½ W	225	48	6	15	42	3 N	28 E	N	8	5	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Precambrian *Sampling:* Statistics are based on 44 specimens from 16 oriented samples.
Bonito Canyon Quartzite	79	36 N	109 W	31	−25	4	19	74	33 N	34 E	S	4	2	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Precambrian *Sampling:* Statistics are based on 74 specimens from 16 oriented samples.
BLACKHEAD SANDSTONE, NEWFOUNDLAND	80	47 N	53 W	232	51	10	25	10	5 N	84 E	—	13½	9	a	*Nairn, Frost, and Light* a Nairn, Frost, and Light, 1959	*Age:* Precambrian *Sampling:* Statistical analysis is based on mean direction of each of 10 oriented samples from one locality. 4 specimen measurements were made on each oriented sample. *Other:* CCW rotation of Newfoundland of 20° is suggested in Ref. a in order to bring these results into closer agreement with results from Hakatai shale; "corrected" pole is then at 10° N 110° W.
	81	47 N	53 W	232	51	10	25	10	2 N	95 W	S	13½	9	*		
SIGNAL HILL SANDSTONE OF NEWFOUNDLAND	82	47 N	53 W	283	20	11½	21	9	16 N	142 W	—	12	6	a	*Nairn, Frost, and Light* a Nairn, Frost, and Light, 1959	*Age:* Precambrian, possibly 8500 feet stratigraphically below Blackhead sandstone *Sampling:* Statistics are based on 9 oriented samples from one locality. 4 specimen measurements were made on each sample. *Stability:* Beds at the Signal Hill site dip steeply eastward, those at the Blackhead site dip less steeply; agreement between the two sites is presumably better after correction for dip. *Other:* Correction for hypothetical rotation of Newfoundland was made as in the case of the Blackhead sandstone. "Corrected" pole is at 29° N, 163° W.
	83	47 N	53 W	283	20	11½	21	9	16½N	145 W	S	12	6	*		
BLACKHEAD AND SIGNAL HILL FORMATIONS, UNDIFFERENTIATED	84	47 N	53 W	262	39	13½	8	19	11 N	122 W	—	16	9½	a	*Nairn, Frost, and Light* a Nairn, Frost, and Light, 1959	*Sampling:* Analysis was made on entire 19 oriented samples of last entries.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION B—*Continued*																
AUSTRALIA																
BULDIVA QUARTZITE	85	14 S	132 E	243	38	12	—	—	30 N	121 W	N	14	8	a	*Irving and Green* a Irving and Green, 1958	*Age:* latest Precambrian; contains fossils of primitive life
NULLAGINE LAVAS	86	21 S	120 E	143	64	8	—	—	51 N	18 W	N	13	10	a	*Irving and Green* a Irving and Green, 1958	*Age:* late Precambrian, part of Catherine River group; probably younger than the Edith River volcanic rocks
EDITH RIVER VOLCANIC ROCKS	87	13 S	132 E	90	48	18	—	—	6 N	14 E	N	24	15	a	*Irving and Green* a Irving and Green, 1958	*Age:* earliest late Precambrian
	88	13 S	132 E	90	48	18	—	—	6 N	14 W	N	24	15	b	b Irving, 1959	
AFRICA																
PILANSBERG DYKES	89	26 S	28 E	24	69½	—	—	—	7½N	42½E	S	—	—	a	*Gough* a Gough, 1956	*Age:* Generally regarded as pre-Karroo, post-Waterberg. Possibility of their being late Precambrian is mentioned in Ref. a. Recently determined radio-isotope age is 1290 m.y. (Schreiner, 1958, p. 1330). *Sampling:* Sampling sites span 54 miles. Vertical sampling extent varied, but usually was at least several hundred feet. Between 8 and 58 samples were collected from each dike studied, giving circles of confidence of 3.3° to 7°. Statistical analysis of entry (90) is based on the mean direction of magnetization of each of 5 dikes. A comparatively small number of randomly magnetized samples is not included in this analysis. *Stability:* Most specimens were stable in alternating fields of the order of 100 oersted.
	90	**26 S**	**28 E**	**24**	**69½**	**6**	**124**	**5**	**7½N**	**42½E**	**S**	**10**	**9**	*		
BUSHVELD GABBRO	91	25½S	28 E	—	—	—	—	5	23 N	36 E	S	—	—	a	*Gough and van Niekerk* a Gough and van Niekerk, 1959	*Age:* Concordant radio-isotope results indicate an age of 2.0×10^9 years. *Sampling:* Between 12 and 29 oriented samples were collected at 5 sites in the Main Gabbro zone of the complex. Lateral sampling extent is 150 miles.
	92	**25½S**	**28 E**	**—**	**—**	**12**	**40**	**5**	**23 N**	**36 E**	**S**	**12**	**12**	*		

Circles of confidence at the 99% level for the directions at each site range between 2° and 9.6°; all data

| | | |

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION C EARLY PALEOZOIC																
DEVONIAN																
EUROPE																
OLD RED SANDSTONE	1	52 N	2½W	233	−22	—	—	39	—	—	—	—	—	a	*Clegg, Almond, and Stubbs*	*Age:* Devonian; Lower part of Old Red sandstone, Brownstone series from Mitcheldean in Gloucestershire
	2	**52 N**	**2½W**	**233**	**−22**	**11½**	**4**	**39**	**31 N**	**111 E**	**N**	**12**	**6½**	*	a Clegg, Almond, and Stubbs, 1954a	*Sampling:* Measurements were made of 39 specimens from 3 oriented samples collected alt one locality. α_{95} (entry 2) was recalculated from vaue of α_{10} given in Ref. a; this is probably based on $N = 39$. The value of k was calculated by the approximate formula.
	3	—	—	—	—	—	—	—	25 N	102 E	N	—	—	b	b Irving, 1959	*Stability:* Magnetizations were stable in D.C. fields of several oersted and in alternating fields of several hundred oersted.
	4	—	—	199	−2	—	—	—	—	—	N	—	—	a	*Creer*	*Age:* Devonian; the Old Red sandstone of the Angleo Welsh cuvette is divided into an upper and lowr-unit; the lower unit consists of the Downtonian and the overlying Dittonian. Entries (6) and (7) refer to the Downtonian, entries (4) and (5) to the two overlying units, and all the other entries refer, presumably, to the entire section of Old Red sandstone.
	5	52 N	3 W	199	−2	—	—	—	37 N	153 E	N	—	—	*	a Creer, Irving, and Runcorn, 1954	*Sampling:* Samples were collected at 17 localities with different amounts of folding. Only the samples from 6 localities with flat-lying strata gave consistent results. The statistical analysis of entries (11), (12), and (13) is based on 35 oriented samples from these 6 localities. The stratigraphic thicknesses sampled at these localities are 30, 80, 250, 0.5, 34, 0.5, and 1200 feet, respectively.
	6	—	—	34	−2	—	—	—	—	—	S	—	—	a	b Runcorn, 1955a	
	7	52 N	3 W	34	−2	—	—	—	31½N	136 E	S	—	—	*	c Runcorn, 1955b	
OLD RED SANDSTONE	8	—	—	—	—	—	—	—	34 N	156 E	M	—	—	a	d Creer, 1957b	
	9	—	—	198	−2	—	—	—	45 N	155 E	N	—	—	b	e Creer, Irving, and Runcorn, 1957	
	10	—	—	—	—	—	—	—	34 N	156 E	N	10	5	c		
	11	52 N	3 W	196	−4	5	19	35	—	—	—	—	—	d		
	12	—	—	196	−5	5	19	35	30 N	159 E	N	5	3	e		
	13	**52 N**	**3 W**	**196**	**−4**	**5**	**19**	**35**	**38½N**	**156 E**	**N**	**5**	**2½**	*		

Stability: The lack of agreement between strata which have been folded by different amounts indicates that at least some of the magnetization is unstable.
Reversals: Ref. d states that reversals are absent from

																6 localities. The stratigraphic thicknesses sampled at these localities are 30, 80, 250, 0.5, 34, 0.5, and 1200 feet, respectively.
																Stability: The lack of agreement between strata which have been folded by different amounts indicates that at least some of the magnetization is unstable. *Reversals:* Ref. d states that reversals are absent from the Old Red sandstone. However, in Ref. a, reversals were noted at two localities in the Downtonian.
NORTH AMERICA																
ONONDAGA LIMESTONE	14	42½N	74 W	177	79	4	19	65	21 N	73 W	S	7½	7	*	*Graham* a Graham, 1956	*Age:* Top of lower or bottom of Early or Middle Devonian (late Ulsterian) *Sampling:* 65 oriented samples were collected at 2 localities. Analysis of entry (14) is based on data scaled from Fig. 3 of Ref. a. *Stability:* Directions of magnetization are approximately parallel throughout both stratiform beds and beds showing penecontemporaneous deformation, indicating that remanent magnetization is post-depositional.
AUSTRALIA																
AINSLIE VOLCANIC ROCKS	15	35 S	149 E	17	−30	12	—	—	66 N	168 W	S	13	7	a	*Irving and Green* a Irving and Green, 1958 b Irving, 1959	*Age:* There is some uncertainty as to whether the volcanic rocks are Devonian or Silurian (Ref. b).
SILURIAN																
EUROPE																
LUDLOW SERIES	16	—	—	205	−16	—	—	—	40 N	140 E	N	—	—	a	*Creer* a Creer, Irving, and Runcorn, 1954	*Age:* The Ludlow series at Pembrokeshire is Late Silurian.
	17	52 N	5 W	205	−16	—	—	—	41 N	141 E	N	—	—	*		
URAL PERIDOTITES	18	67 N	66 E	98	38	—	—	—	16 N	140 E	S	—	—	a	*Komarov* a Komarov, 1959	*Age:* Silurian *Sampling:* 6 samples were collected at one quarry.
NORTH AMERICA																
ROSE HILL FORMATION OF SWARTZ, 1923 (Sediments)	19	40 N	78 W	322	−39	5	—	—	19 N	138 E	S	6	4	b	*Graham* a Graham, 1949 b Irving, 1956a c Creer, Irving, and Runcorn, 1957	*Age:* early Middle Silurian (lower Niagara series) *Sampling:* 35 oriented samples were collected at 6 localities spanning a distance of 32 miles. Entry (21) was based on data scaled from Figs. 18, 20, and 22 of Ref. a. *Stability:* Samples were collected on limbs of small
	20	—	—	322	−39	5	27	—	19 N	138 E	S	6	4	c		
	21	35½N	79 W	321	−40	6½	16	35	18 N	138½E	S	8	5	*		

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION C—*Continued*																
																folds at 2 localities and on steep limbs of large folds at the remaining localities. Directions of magnetization are widely scattered before correction for dip. All but 3 samples have nearly parallel directions after dip correction. Deformation was near the end of the Paleozoic, indicating magnetic stability for at least 200 million years. *Reversals:* Of the 3 aberrant samples, 2 that are stratigraphically adjacent have directions of magnetization approximately opposed to the mean direction for the entire group.
CLINTON IRON ORE	22	—	—	—	—	—	—	—	35 N	138 E	N	—	—	a	*Howell et al.* a Howell, Martinez, and Statham, 1958	*Age:* Middle Siluvian (Niagaran series) *Sampling:* 16 specimens were measured from 7 oriented samples. Lateral sampling extent: 300 feet. Analysis (entry 23) is based on data scaled from Fig. 1 of Ref. a. Since specimen rather than sample directions are used, α_{95} may be too small. *Other:* Planes of maximum susceptibility tend to lie in bedding plane.
	23	**33½N**	**86½W**	**143**	**19½**	**11½**	**107**	**16**	**34 N**	**139 E**	**N**	**12**	**6½**	*		
AUSTRALIA																
MUGA PORPHYRY	24	35 S	149 E	26	−30	22	—	—	60 N	157 W	S	24	14	a	*Irving and Green* a Irving and Green, 1958	*Age:* Late Silurian (Ref. a)
	25	**35 S**	**149 E**	**26**	**−30**	**22**	**—**	**—**	**60 N**	**153 W**	**S**	**24**	**14**	*		
ASIA																
RED SILTSTONES FROM YUMEN	26	—	—	293½	55½	8½	—	17	49 N	12 E	S	—	—	a	*Chang Wen-You and Nairn* a Chang Wen-you and Nairn, 1959	*Age:* Middle Silurian *Sampling:* Analysis of Ref. a is based on measurements of 17 specimens from 3 oriented samples. Collecting site is described as southern part of Yumen, Kansu province.
	27	**40 N**	**97 E**	**293½**	**55½**	**8½**	**16**	**17**	**38½N**	**25½E**	**S**	**12**	**8½**	*		
ORDOVICIAN																
EUROPE																

															a Chang Wen-you and Nairn, 1959	of 17 specimens from 3 oriented samples. Collecting site is described as southern part of Yumen, Kansu province.

ORDOVICIAN																
EUROPE																
	28	51 N	26 E	140	75	—	—	—	28 N	46 E	S	—	—	a	*Komarov*	*Age:* Shown as Ordovician(?) in Table 2 of Ref. a, but Komarov (1959) believes European polar-wandering curve suggests Cambrian age for entries (28, 29) and Ordovician age for entries (30, 31).
UKRAINIAN BASALTS	**29**	**51 N**	**26 E**	**140**	**74½**	**10½**	**40**	**6**	**27½ N**	**46 E**	**S**	**19½**	**17½**	*	a Komarov, 1959	*Sampling:* 6 samples collected in a quarry were reported for entries (28) and (29), and 8 samples from 2 quarries for entries (30) and (31). Data for calculating entries (29) and (31) were taken from Table 1, Ref. a.
	30	51 N	26 E	255	58	—	—	—	21 N	27 W	S	—	—	a		
	31	**51 N**	**26 E**	**255**	**56½**	**11½**	**23**	**8**	**20 N**	**29 W**	**S**	**16½**	**12**	*		
	32	60 N	30 E	211	−35	—	—	—	—	—	N	—	—	a	*Khramov*	*Age:* Early Ordovician
RED SANDS AND BROWN CLAYS NEAR LENINGRAD	33	60 N	30 E	27	58	—	—	—	—	—	S	—	—	a	a Khramov, 1958	*Sampling:* 29 oriented samples were collected over a stratigraphic thickness of 10 meters. Entry (34) is an estimate that excludes an unstable component in entry (33). Entry (35) is from Table 25, Ref. a, and entry (36) is the mean of the poles in entries (32) and (34).
	34	60 N	30 E	38	41	—	—	—	—	—	S	—	—	a		
	35	60 N	30 E	—	—	18½	—	—	42 N	169 E	M	—	—	a		
	36	**60 N**	**30 E**	**—**	**—**	**18½**	**—**	**—**	**44 N**	**162 E**	**M**	**22**	**13**	*		
NORTH AMERICA																
TRENTON GROUP (sediments)																
Sprakers, New York	**37**	**42½N**	**75 W**	**177½**	**71**	**5½**	**23**	**28**	**8½N**	**74 W**	**S**	**9½**	**8**	*	*Graham* a Graham, 1956	*Age:* Middle Ordovician *Sampling:* 28 oriented samples were collected, presumably at 1 locality; samples are from different cobbles in a limestone conglomerate. Data for entry (37) are based on values scaled from Fig. 2 of Ref. a. *Stability:* Agreement in directions of magnetization in different cobbles indicates that the remanent magnetization is not that of the parent limestone body but was acquired after the deposition of the Trenton group.
Trenton Falls, New York	**38**	**43½N**	**75 W**	**179**	**82**	**5**	**23**	**45**	**27½N**	**75 W**	**S**	**10**	**9½**	*	b Graham, 1954	*Sampling:* Data for entry (38) were scaled from Fig. 5 of Ref. b showing 35 measurements from flat-lying beds and 10 from a local deformed bed after correcting for dip. *Stability:* Flat-lying beds have well-grouped directions of magnetization. A distorted zone enclosed by flat-lying beds has scattered directions, which tend to move toward the other group on correcting for dip; significant stability since deposition is thus indicated (Graham, 1954, p. 219).

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION C—*Continued*																
JUNIATA FORMATION (sediments)	**40**	**40 N**	**78½W**	**131**	**26**	**8**	**6**	**56**	**20 N**	**153 E**	N	**9**	**5**	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Late Ordovician *Sampling:* Statistical analysis is based on 56 specimens from 12 oriented samples.
CAMBRIAN																
EUROPE																
CAERBWDY SANDSTONE	41	—	—	187	39	—	—	—	15 N	173 E	N	—	—	a	*Creer* a Creer, Irving, and Runcorn, 1954 b Runcorn, 1955a c Irving, 1956a d Creer, Irving, and Runcorn, 1957 e Creer, 1957b	*Age:* Cambrian. Caerbwdy sandstone is in the lower part of the Cambrian Caerfai series. *Sampling:* 12 samples span a stratigraphic interval of 350 feet. *Other:* Values cited in entry (41) also appear in Ref. b. Details appear in Ref. e, p. 123–124.
	42	—	—	191	41	—	—	—	15 N	170 E	N	—	—	b		
	43	52 N	5 W	187	39	8	—	—	15 N	173 E	—	10	7	c		
	44	—	—	187	39	8	32	12	15 N	173 E	N	10	8	d		
	45	**52 N**	**5 W**	**187**	**39**	**8**	**32**	**12**	**15½N**	**168½ E**	N	**9½**	**5½**	*		
NORTH AMERICA																
TAPEATS SANDSTONE, GRAND CANYON	46	—	—	—	—	—	—	—	22 N	27 E	—	—	—	a	*Runcorn* a Day and Runcorn, 1955 b Creer, Irving, and Runcorn, 1957	*Age:* Early Cambrian *Other:* No other data about this research are available. According to Ref. b, the pole of Ref. a is based on inadequate data and should be disregarded.
WILBERNS FORMATION (sediments)	**47**	**30½N**	**99 W**	**98**	**24½**	**—**	**—**	**—**	**0**	**158 E**	N	**—**	**—**	a	*Howell and Martinez* a Howell and Martinez, 1957	*Age:* Late Cambrian; Point Peak shale member of Wilberns formation, Llano uplift area *Sampling:* 185 samples were collected at 10 localities spanning a distance of 55 miles. Statistical analysis was not made of this data because directions of magnetization show "streaking" toward direction of present field. Values for declination and inclination indicated are for the group of measurements farthest from the present field direction; they were scaled from Fig. 5, Ref. a, and do not exactly correspond
																Stability: "Streaking" indicates partial instability. A single locality with steeply dipping beds shows wide

present field. Values for declination and inclination indicated are for the group of measurements farthest from the present field direction; they were scaled from Fig. 5, Ref. a, and do not exactly correspond

																Stability: "Streaking" indicates partial instability. A single locality with steeply dipping beds shows wide scatter in directions of magnetization after correcting for dip. *Reversals:* No systematic reversals occur, but some widely scattered points are on the upper hemisphere.
SAWATCH "QUARTZITE" SANDY DOLOMITE	48	—	—	—	—	—	—	—	49 N	125 E	N	—	—	b	*Howell and Martinez*	*Age:* Late Cambrian; magnetization may be post depositional
	49	**39 N**	**106½W**	**148**	**−15**	**4**	**44**	**31**	**47 N**	**125 E**	**N**	**4**	**2**	*	a Howell and Martinez, 1957 b Howell, Martinez, and Statham, 1958	*Sampling:* 36 samples were collected at 2 localities. Analysis for entry (49) was based on data scaled from Fig. 7 of Ref. a, omitting 5 samples with directions parallel to present field. *Stability:* 5 of the 36 samples have directions of magnetization parallel to the present field; 31 have tight grouping approximately perpendicular to the present field. *Other:* Remanent magnetization may have been acquired at time of dolomitization, possibly in the late Paleozoic, Ref. a, p. 391.
LODORE FORMATION (sediments)	50	41 N	109½ W	59	4	8	14	26	—	—	—	—	—	a	*Collinson and Runcorn*	*Age:* Cambrian *Sampling:* Based on 33 specimens from 11 oriented samples. *Reversals:* Entry (50) is reversed with respect to (51). Stratigraphic distribution of samples in these two groups is not known.
	51	41 N	109½ W	234	13	13	25	7	—	—	—	—	—	a		
	52	**41 N**	**109½ W**	**—**	**—**	**—**	**—**	**—**	**23 N**	**6 E**	**M**	**7**	**4**	a	a Collinson and Runcorn, 1960	
DEADWOOD FORMATION (sediments)	**53**	**42 N**	**107½ W**	**151**	**−14**	**7**	**15**	**34**	**47 N**	**117 E**	**N**	**7**	**4**	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Cambrian. Ref. a cites a possible Mississippian age. However, it is definitely regarded as Cambrian by U. S. Geological Survey. *Sampling:* 34 specimen measurements were made on 7 oriented samples.
AUSTRALIA																
ELDER MOUNTAIN SANDSTONE	54	16 S	126 E	231	−15	10	—	—	34 N	172 W	N	10	5	a	*Irving and Green*	*Age:* Middle Cambrian (Ref. a)
	55	**16 S**	**126 E**	**231**	**−15**	**10**	**—**	**—**	**34 N**	**165 W**	**N**	**10**	**5**	*	a Irving and Green, 1958	
ANTRIM PLATEAU BASALTS	**56**	**16 S**	**126 E**	**53**	**−2**	**12**	**—**	**—**	**36 N**	**154 W**	**M**	**12**	**6**	a	*Irving and Green* a Irving and Green, 1958 b Irving, 1959	*Age:* Early Cambrian, possibly latest Precambrian (Ref. b) *Reversals:* Reversals are indicated in Ref. b, but no details are given.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION D CARBONIFEROUS																
EUROPE																
GLOUCHESTER	1	—	—	33	35	—	—	14	—	—	S	—	—	a	*Clegg et al.*	*Age:* late Carboniferous (late Coal Measures)
PENNANT	**2**	**51½N**	**2½W**	**33**	**35**	**10½**	**13**	**14**	**48½N**	**126½E**	**S**	**12**	**6½**	*	a Clegg, Almond, and Stubbs, 1954a	*Sampling:* 14 specimen measurements were made on 1 oriented sample. Calculations for α_{95} and k in entry (2) were based on $\alpha_{50} = 5$, cited in Ref. a.
SANDSTONE	3	—	—	—	—	—	—	—	43 N	114 E	S	—	—	b	b Irving, 1959	*Stability:* The magnetization of the specimens was unchanged after application of fields of several hundred oersted A.C.
CLEE HILL SEDIMENTS AND	4	—	—	200	15	—	—	45	—	—	N	—	—	a	*Clegg et al.*	*Age:* Probably late Carboniferous
IGNEOUS ROCKS	**5**	**52½N**	**2 W**	**200**	**15**	**3½**	**36**	**45**	**27½N**	**156 E**	**N**	**3½**	**2**	*	a Clegg, Deutsch, Everitt, and Stubbs, 1957	*Sampling:* Samples were collected at 6 sites over a lateral extent of 30 miles. It is not known whether N is the number of oriented samples or specimen measurements. Calculations for entry (5) were based on data scaled from Fig. 1, Ref. a. Igneous and sedimentary samples could not be distinguished in this plot.
	6	—	—	—	—	—	—	—	27 N	155 E	N	—	—	b	b Irving, 1959	*Stability:* Stability is suggested by the tight grouping of the baked sediments compared with scattered directions and lower intensity of nearby unbaked sediments.
TIDESWELLDALE	7	—	—	218	36	—	—	5	—	—	—	—	—	a	*Clegg et al.*	*Age:* These sediments were baked by an intrusive of late Carboniferous age.
BAKED SEDIMENTS	**8**	**53½N**	**2 W**	**219**	**41**	**8**	**61**	**5**	**6½N**	**142½E**	**N**	**9½**	**6**	*	a Clegg, Deutsch, Everitt, and Stubbs, 1957	*Sampling:* It is not known whether N is the number of oriented samples or specimen measurements. Calculations for entry (8) were based on data scaled from Fig. 1, Ref. a.
	9	—	—	—	—	—	—	—	5 N	143 E	N	—	—	b	b Irving, 1959	
LANCASHIRE PENDLE MONOCLINE	**10**	**54 N**	**3 W**	**24½**	**23½**	**4**	**64**	**19**	**44 N**	**142½ E**	**S**	**4**	**2**	*	*Belshé* a Belshé, 1957	*Age:* Carboniferous *Sampling:* 19 oriented samples were collected at 3 sites from the lower Coal Measures, Dandy Rock and Old Lawrence Rock. Calculations for entry (10) were based on data scaled from Fig. 3, Ref. a. *Stability:* Application of Graham's fold test indicates that the magnetization is stable and was imparted to the rocks before late Carboniferous folding.
LANCASHIRE ROCKS	**11**	**54 N**	**3 W**	**27**	**24**	**6**	**29**	**18**	**43 N**	**139 E**	**S**	**6½**	**3½**	*	*Belshé* a Belshé, 1957	*Age:* late(?) Carboniferous *Sampling:* Calculations for entry (11) were based on data scaled from Fig. 2, Ref. a. This figure shows

LANCASHIRE ROCKS	11	54 N	3 W	27	24	6	29	18	43 N	139 E	S	6½	3½	°	*Belshé* a Belshé, 1957	*Age:* late(?) Carboniferous *Sampling:* Calculations for entry (11) were based on data scaled from Fig. 2, Ref. a. This figure shows the directions of magnetization for 40 (out of a reported 49) samples. These 18 samples form 1 group and the 13 used for the following entry (12) form another group. 9 scattered directions were excluded from these calculations.
LANCASHIRE MILLSTONE GRIT	12	54 N	3 W	188½	9½	3	176	13	30½N	167 E	N	3	1½	°	*Belshé* a Belshé, 1957	*Age:* early late Carboniferous *Sampling:* Calculations for entry (12) were based on data scaled from Fig. 2, Ref. a. Also see remarks for entry (11) above.
DERBYSHIRE SANDSTONE AND SILTSTONE	13	—	—	26	37	—	—	—	—	—	S	—	—	a	*Belshé* a Runcorn, 1955a	*Age:* early and late Carboniferous *Sampling:* 103 oriented samples were collected from 14 sites. *Stability:* Stability is suggested by remeasurement after 6 months with no change in the magnetization. *Other:* A later study cited in Ref. c of 142 samples from 34 localities showed much greater scatter.
	14	—	—	—	—	—	—	—	36 N	137 E	S	—	—	b	b Runcorn, 1955b	
	15	—	—	27	36	—	—	103	—	—	S	—	—	c	c Belshé, 1957	
	16	**53 N**	**1½W**	**27**	**36**	**—**	**—**	**103**	**50½N**	**136 E**	**S**	**—**	**—**	°		
	17	53 N	1½W	26	37	—	—	—	51 N	143 E	S	—	—	d	d Irving, 1956a	
DERBYSHIRE TOADSTONES	18	53 N	1½W	26	43	—	—	—	55 N	148 E	S	—	—	a	*Belshé* a Irving, 1956a	*Age:* The lavas are interbedded with sediments of early Carboniferous age and are probably pre-Millstone Grit (Evans, 1918, p. 172). *Sampling:* 9 oriented samples were obtained from 3 interbedded units. Calculations in entry (20) were based on data scaled from Fig. 1, Ref. b.
	19	—	—	48	47	13	—	9	—	—	—	—	—	b	b Belshé, 1957	
	20	**53 N**	**1½W**	**48**	**47**	**13**	**13**	**9**	**47 N**	**105 E**	**S**	**17**	**11**	°		
KINGHORN LAVAS															*Clegg et al.*	*Age:* early Carboniferous (pre-Millstone Grit)
Flows 64–65	21	56 N	3½W	20	15	—	—	—	—	—	—	—	—	a	a Clegg, Deutsch, Everitt, and Stubbs, 1957 b Irving, 1959	*Sampling:* The sequence of flows is as indicated, with 65 uppermost. Calculations for entries (27) and (29) (flows 50–54 and 41–46) were based on data scaled from Fig. 2, Ref. a. Data for flows 64–65 and 48–49 were not given. Entry (31) is a Fisher statistical treatment of the pole positions given in entries (22), (25), and (27).
	22	56 N	3½W	20	15	—	—	—	39 N	150½E	S	—	—	°		
	23	—	—	—	—	—	—	—	42 N	150 E	S	—	—	b		
Flows 48–54	24	56 N	3½W	200	38	—	—	—	—	—	—	—	—	a		
	25	56 N	3½W	200	38	—	—	—	10½N	157½E	N	—	—	°		
	26	—	—	—	—	—	—	—	12 N	157½E	N	—	—	b		
Flows 50–54	27	56 N	3½W	202½	34	5½	23	27	13 N	154½E	N	6½	3½	°		
Flows 41–46	28	56 N	3½W	26	−42	—	—	18	—	—	—	—	—	a		
	29	56 N	3½W	26	−42	3	156	18	6½N	153 E	S	4	2½	°		
	30	—	—	—	—	—	—	—	8 N	153 E	S	—	—	b		
Kinghorn Average	**31**	**—**	**—**	**—**	**—**	**28**	**21**	**3**	**18½N**	**154 E**	**M**	**28**	**28**	°		

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION D—*Continued*																
SHATTERFORD INTRUSION	32	—	—	—	—	—	—	—	32 N	137 E	M	—	—	b	*Clegg et al.*	*Age:* Cited as Carboniferous (undifferentiated) in Ref. b
	33	**52½N**	**2 W**	**31½**	**3**	**4**	**63**	**21**	**32½N**	**139½E**	**M**	**4**	**2**	*	a Clegg, Deutsch, Everitt, and Stubbs, 1957 b Irving, 1959	*Sampling:* The samples were collected at 2 sites 200 yards apart; it is not known whether N is the number of oriented samples or specimen measurements. Calculations in entry (33) were based on data scaled from Fig. 3, Ref. a. *Reversals:* 11 of the 21 samples were oppositely polarized from the other 10.
LUNDY GRANITES	34	—	—	175	−9	27	—	5	—	—	—	—	—	a	*Blundell*	*Age:* Cited as Permian and Carboniferous in Ref. a and as late Carboniferous in Ref. b.
	35	—	—	—	—	—	—	—	43 N	180 E	N	—	—	b	a Blundell, 1957	*Sampling:* 10 specimen measurements were made on 5 oriented samples. The value of k for entry (36) was calculated by the approximate formula.
	36	**51 N**	**4½W**	**175**	**−9**	**27**	**5**	**5**	**43 N**	**177 W**	**N**	**27**	**13½**	*	b Irving, 1959	
NORTH AMERICA																
NACO FORMATION															*Runcorn*	*Age:* Pennsylvanian
Carizzo Creek (sediments)	37	36 N	113 W	149½	−3½	4	40	31	48½N	120 E	N	4	4	a	a Runcorn, 1956a	*Sampling:* Entries (37)–(39) are based on measurements of 31 specimens from 8 oriented samples collected at 1 site with a stratigraphic extent of 200 feet. Entry (40) is from a later collection of 9 oriented samples made at Fossil Creek.
	38	**36 N**	**113 W**	**149½**	**−3½**	**4**	**40**	**31**	**45½N**	**114 E**	**N**	**4**	**2**	*	b Irving, 1959	
	39	—	—	—	—	—	—	—	41 N	120 E	N	8	4	b		
Fossil Creek (limestone)	**40**	**34½ N**	**111½ W**	**125**	**16**	**7**	**22**	**20**	**23 N**	**130 E**	**N**	**7**	**4**	**a**	*Collinson and Runcorn* a Collinson and Runcorn, 1960	
BARNETT FORMATION (sediments)	41	31 N	99 W	—	—	—	—	—	39 N	124 E	N	—	—	a	*Howell and Martinez*	*Age:* Mississippian
	42	31 N	99 W	—	—	—	—	—	41 N	128 E	N	—	—	b		*Sampling:* 60 oriented samples from 8 sites (N poles) and 8 oriented samples from 1 site (S poles) were collected over a lateral extent of 73 miles. Calculations for entries (43) and (45) were based on data scaled from Fig. 3, Ref. b.
	43	**31 N**	**99 W**	**148½**	**19**	**5½**	**11**	**60**	**39 N**	**122½E**	**N**	**6½**	**3**	*	a Martinez and Howell, 1956	
	44	31 N	99 W	—	—	—	—	—	42 N	142 E	S	—	—	a		
	45	**31 N**	**99 W**	**319**	**8**	**3½**	**200**	**8**	**42½N**	**144 E**	**S**	**3½**	**2**	*	b Howell and Martinez, 1957	
CODROY GROUP (sediments)	46	48 N	59 W	166	8	8½	38	9	43 N	139 E	N	8½	4½	a	*Nairn et al.*	*Age:* Cited as Mississippian in Ref. a
	47	**48 N**	**59 W**	**166**	**8**	**8½**	**38**	**9**	**36½N**	**139 E**	**N**	**8½**	**4½**	*	a Nairn, Frost, and Light, 1959	*Sampling:* 36 specimen measurements were made on

	45	31 N	99 W	319	8	3½	200	8	42½N	144 E	S	3½	2	*	b Howell and Martinez, 1957	lations for entries (43) and (45) were based on data scaled from Fig. 3, Ref. b.
CODROY GROUP (sediments)	46	48 N	59 W	166	8	8½	38	9	43 N	139 E	N	8½	4½	a	*Nairn et al.*	*Age:* Cited as Mississippian in Ref. a
	47	48 N	59 W	166	8	8½	38	9	36½N	139 E	N	8½	4½	*	a Nairn, Frost, and Light, 1959	*Sampling:* 36 specimen measurements were made on 9 oriented samples.
BONAVENTURE, KENNEBECASIS and BATHURST FORMATIONS (sediments)	48	48 N	66 W	163½	19½	5	—	46	—	—	—	—	—	a	*Du Bois*	*Age:* Carboniferous; Bonaventure is Late Mississippian or Early Pennsylvanian, Kennebecasis is Pennsylvanian (Ref. a)
	49	48 N	66 W	163½	19½	5	17	46	30 N	133 E	N	5	2½	*	a Du Bois, 1959b	*Sampling:* 22, 14, and 10 oriented samples were collected from these 3 formations from groups of sites spanning 250 miles. The mean direction of 2 specimen measurements from each sample was reduced to the common mean locality cited, and a statistical analysis was made on the resulting directions. α_{95} may be too low because mean site or formation directions were not used in the analysis. The value of k for entry (49) was calculated by the approximate formula.
AUSTRALIA																
KATTUNG VARVOID SEDIMENTS	50	33 S	151 E	90	84	6	—	75	32 N	15 W	M	12	12	a	*Irving*	*Age:* Age cited as late Carboniferous in Ref. b
															a Irving, 1957b b Irving and Green, 1958	*Sampling:* 75 oriented samples were collected at 4 localities. *Stability:* Application of Graham's fold test indicates stability. *Reversals:* The direction of magnetization of samples from one locality was reversed to that at the other 3 localities.
KATTUNG LAVAS	51	33 S	151 E	5	−85	8	—	—	43 N	30 W	S	16	16	a	*Irving and Green* a Irving and Green, 1958	*Age:* Cited as late Carboniferous in Ref. a
AFRICA																
DWYKA VARVED CLAYS	52	18 S	29 E	360	−81	5½	84	10	36 N	151 W	S	10	10	a	*Nairn*	*Age:* Cited as late Carboniferous in Ref. a
	53	18 S	29 E	333	76	7	57	9	7 N	17 E	—	13	12	a	a Creer, Irving, Nairn, and Runcorn, 1958	*Sampling:* The 19 specimen measurements (including "normal" and "reversed" groups) were made on 4 oriented samples.
	54	18 S	29 E	333	76	7	57	9	5½N	17½E	S	13	12	*		

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION E PERMIAN																
EUROPE																
Exeter Volcanic Series	1	—	—	189	−9	—	—	—	47 N	147 E	N	—	—	a	*Creer*	*Age:* Cited as Permian
	2	—	—	—	—	—	—	—	48 N	168 E	N	—	—	c	a Creer, Irving, and Runcorn, 1954	*Sampling:* 34 oriented samples were collected from 5 flows. Pole position cited from Ref. a was scaled from Fig. 1. This figure also appears in Ref. b. Values cited for Ref. c were also scaled from a map. Data for the individual flow directions are given in Ref. e. The dispersion within flows is much less than that between flows.
	3	—	—	189	−9	20	15	5	43 N	164 E	N	20	10	f	b Runcorn, 1955a	
	4	**51 N**	**4 W**	**189**	**−9**	**20**	**—**	**5**	**43 N**	**164 E**	**N**	**20**	**10**	**d**	c Runcorn, 1956a	
															d Irving, 1956a	
															e Creer, 1957b	
															f Creer, Irving, and Runcorn, 1957	
Mauchline Sediments	5	—	—	187	−6	12	—	26	37 N	163 E	N	—	—	a	*Du Bois*	*Age:* Cited as Permian
	6	**55½N**	**4½W**	**187**	**−6**	**12**	**5**	**26**	**37 N**	**166½E**	**N**	**12**	**6**	*	a Du Bois, 1957	*Sampling:* 26 is the number of specimen measuree ments; number of samples is not known. The valu³ of k for entry (6) was calculated by the approximat- formula.
Mauchline Lavas	7	—	—	180	−4	8	—	34	36 N	175 E	N	—	—	a	*Du Bois*	*Age:* Cited as Permian
	8	**55½N**	**4½W**	**180**	**−4**	**8**	**9**	**34**	**36 N**	**175 E**	**N**	**8**	**4**	*	a Du Bois, 1957	*Sampling:* 34 is the number of specimen measure- ments. The number of flows sampled is not given. The value of k for entry (8) was calculated by the approximate formula.
Esterel Volcanic Rocks																*Age:* Cited as Permian
Pyromeride R4	9	—	—	210	−16	—	—	5	—	—	—	—	—	a	*Roche*	*Sampling:* The samples were collected "several dozen meters apart and from different levels." The value of α_{95} for entry (10) is a minimum value which assumes that the 5 samples all came from different flows and is based on the "secular variation precision" $k' = 30$. The value of α_{95} for entry (13) is based on the sample data cited on page 2953, Ref. a, and is greater than
	10	**43½N**	**7 E**	**210**	**−16**	**11½**	**k'**	**5**	**46 N**	**142 E**	**N**	**12**	**6½**	*	a Roche, 1957	
	11	—	—	—	—	—	—	—	46 N	141 E	N	—	—	b	b Irving. 1959	
Dolerite	12	—	—	175	−13	—	—	3	—	—	—	—	—	a		
	13	**43½N**	**7 E**	**175**	**−13**	**18**	**20**	**3**	**52½N**	**165°W**	**N**	**18**	**9**	*		
	14	—	—	—	—	—	—	—	53 N	165°W	N	—	—	b		

the minimum value based on the "secular variation precision."

Stability: The magnetization of the samples was not

Dolerite	13	43½N	7 E	175	−13	18	20	3	52½N	165°W	N	18	9	*		
	14	—	—	—	—	—	—	—	53 N	165°W	N	—	—	b		
Rhyolite	15	43½N	7 E	217	−22½	4½	—	14	45 N	130½E	N	—	—	a	*Rutten et al.*	
	16	43½N	7 E	217	−22½	4½	69	14	45 N	130½E	N	4½	2½	*	a Rutten, van Everdingen, and Zijderveld, 1957	
Undifferentiated	17	—	—	—	—	—	—	14	47 N	144 E	—	—	—	a	*As and Zijderveld*	
	18	43½N	7 E	207½	−16	5	59	14	47 N	145 E	N	5½	3	*	a As and Zijderveld, 1958	
Oslo Graben Trachyandesite	19	60 N	10½E	201½	−33	9½	—	12	45½N	165 E	N	—	—	a	*Rutten et al.*	
	20	60 N	10½E	201½	−33	9½	18	12	45 N	160½E	N	11	6	*	a Rutten, van Everdingen, and Zijderveld, 1957	
Ayrshire Kylites	21	54 N	4½W	181	7	—	—	7	—	—	N	—	—	a	*Armstrong*	
	22	54 N	4½W	190	1½	12	5	5	34 N	163½E	N	12	6	*	a Armstrong, 1957	
Niedeck Porphyre	23	—	—	193	−7	4½	22	49	43 N	168 E	N	—	—	a	*Nairn*	
	24	—	—	193	−7	4½	22	49	43 N	168 E	N	4½	2½	*	a Nairn, 1957b	
Montcenis Sediments	25	—	—	197	6	4	93	14	38 N	162 E	N	—	—	a	*Nairn*	
	26	46½N	4½E	197	6	4	93	14	38 N	162 E	N	4	2	*	a Nairn 1957b	

is based on the "secular variation precision" $k' = 30$. The value of α_{95} for entry (13) is based on the sample data cited on page 2953, Ref. a, and is greater than the minimum value based on the "secular variation precision."

Stability: The magnetization of the samples was not changed by heating to 300°C. and cooling in zero applied field.

Sampling: The 14 samples were all collected from a single flow. Since only one point in time was sampled the circle of confidence does not have the usual significance. The value of k for entry (16) was calculated by the approximate formula.

Sampling: These samples contain dolerites, rhyolites, pelites, and arkoses. Statistical analysis for entry (18) is based on data scaled from Fig. 6(d), Ref. a.

Stability: Partial demagnetization at 150°C. and 300 oersted A.C. field decreases scatter. Application of Graham's fold test also indicates stability.

Sampling: 3 flows and 12 "rhomboporphyries" were sampled over a lateral extent of 30 by 15 km and a stratigraphic thickness of 750 m. The value of k for entry (20) was calculated by the approximate formula.

Stability: Stability was checked by inverting the samples in the laboratory and remeasuring. 12 stable samples were used for the calculations.

Age: The Kylites intrude Coal Measures and are cut by Permian volcanic necks. They are suggested as equivalent to the Mauchline Lavas in Ref. a.

Sampling: Entry (21) is based on Kylite samples from 5 localities plus 2 other Permian localities. Entry (22) is based on the data for the Kylite samples only.

Stability: Partial demagnetization was used to decrease the scatter.

Age: Saxonian(?), middle Permian

Sampling: 49 specimen measurements were made on 14 oriented samples.

Other: Locality is in the Vosges of France; exact locality could not be determined.

Age: Cited as Saxonian, middle Permian

Sampling: 14 specimen measurements were made on 3 oriented samples. Some specimens were not included in the statistical analysis because they were believed to be unstable.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION E—*Continued*																
SAINT-WENDEL SEDIMENTS	27	—	—	181	−9	3½	27	27	45 N	175 W	N	—	—	a	*Nairn* a Nairn, 1957b	*Age:* Cited as Autunian, early Permian *Sampling:* 27 specimen measurements were made on 5 oriented samples.
	28	**49½N**	**7 E**	**181**	**−9**	**3½**	**27**	**27**	**45 N**	**175 W**	N	**3½**	**2**	*		
UFIMSKIJ AND KAZANSKIJ SEDIMENTS	29	58 N	56 E	221	−40	15½	—	—	45 N	178 E	N	—	—	a	*Khramov* a Khramov, 1958	*Age:* late Permian *Sampling:* 16 oriented samples were collected from Ufimskij over a lateral extent of 100 km and a thickness of 70 m, and 24 were collected from Kazanskij over a lateral extent of 100 km and a thickness of 90 m. Data for entry (29) are an estimate which excludes "partially" stable samples. *Other:* A large amount of scatter is present in the plotted measurements.
	30	**58 N**	**56 E**	**221**	**−40**	**15½**	**—**	**—**	**45 N**	**178 E**	N	**18½**	**11½**	*		
TARTARSKIJ SEDIMENTS	**31**	**59 N**	**50 E**	**—**	**—**	**10**	**—**	**—**	**52 N**	**176 E**	**17**	**—**	**—**	a	*Khramov* a Khramov, 1958	*Age:* late Permian *Sampling:* 74 oriented samples were collected over a lateral extent of 100 km and a thickness of 485 m. Khramov states that two groups are present, centering at 38 E, plus 57 and 254 E, plus 42. He states that the latter group has a stable component at 211 E, minus 38. *Other:* A plot of some of these data shows very extreme scatter.
NORTH AMERICA																
CUTLER FORMATION (sediments)															*Graham* a Graham, 1955	*Age:* Permian
Glenwood Springs, Colo.	32	39½N	107½W	—	—	—	—	2	36 N	122 E	N	—	—	a		*Sampling:* 2 samples were collected 50 feet apart laterally and 1 foot apart stratigraphically. Entry (32) values were scaled from a map on p. 343, Ref. a. Entry (33) is based on data scaled from Fig. 7, Ref. a. ($N = 2$ is not sufficient to calculate α_{95} or k.)
	33	39½N	107½W	140	6	—	—	2	33½N	123 E	N	—	—	*		
Monument Valley, Utah	34	37 N	110 W	161	32½	9½	96	4	33 N	92 E	N	11	6	*		*Sampling:* 12 samples were collected over a stratigraphic thickness of 200 feet and a lateral extent of ¼ mile. Entry (34) is based on data scaled from Fig. 7, Ref. a, for 4 samples, which form a group away from the present field direction.
CUTLER, AVERAGE	**35**	—	—	—	—	**13**	—	**2**	**34 N**	**107 E**	N	**13**	**13**	*		*Sampling:* This pole is midway between the poles of entries (33) and (34); 13° is the distance to either pole and has no statistical significance.

Monument Valley, Utah	34	37 N	110 W	155	10	7½	—	—	37 N	110 E	N	—	—	a		graphic thickness of 200 feet and a lateral extent of ¼ mile. Entry (34) is based on data scaled from Fig. 7, Ref. a, for 4 samples, which form a group away from the remainder of the directions.
CUTLER, AVERAGE	35	—	—	—	—	12	—	2	34 N	107 E	N	13	12	•		*Sampling:* This pole is midway between the poles of entries (33) and (34); 13° is the distance to either pole and has no statistical significance.
SUPAI FORMATION																
Upper Supai Oak Creek, Arizona	37	35 N	112 W	161	10	8	13	32	46 N	96 E	N	8	4	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Permian and Pennsylvanian *Sampling:* Statistical analysis is based on 32 specimen measurements on 14 oriented samples.
Oak Creek, Arizona	38	35 N	111½W	—	—	—	—	—	43 N	122 E	N	—	—	a	*Graham*	*Sampling:* 52 specimen measurements were made on 17 oriented samples spanning 1.4 miles laterally and 100 feet stratigraphically. Specimens from conglomerates had rather scattered directions. 11 specimens from flat-lying sediments show streaking toward the present field. The analysis of entry (39) is based on the remaining 24 specimens from flat-lying sediments. Entry (38) was scaled from Fig. 6 of Ref. a.
	39	35 N	111½W	143½	9½	7½	17	24	37½N	117 E	N	7½	4	•	a Graham, 1955 b Irving, 1956a c Irving, 1959	
Carizzo Creek, Arizona	40	34 N	110½W	—	—	—	—	—	50 N	104 E	N	—	—	a		*Sampling:* Statistical analysis of entry (41) is based on 59 specimen measurements made on 30 oriented samples collected over a lateral distance of 10 miles and a stratigraphic thickness of 1000 feet. Entry (40) was scaled from Fig. 6 of Ref. a.
	41	34 N	110½W	159	2	3	33	59	50 N	103 E	N	3	1½	•		
Supai Combined	42	35 N	104 W	150	3	5	—	—	37 N	107 E	—	5	3	b		*Sampling:* Described in Ref. b as Supai Beds, Arizona and New Mexico.
	43	—	—	—	—	—	—	—	41 N	117 E	N	—	—	c		
Grand Canyon and Colorado Plateau	44	—	—	—	—	—	—	—	23 N	119 E	N	8	6	a	*Runcorn*	*Sampling:* 31 samples were collected over a lateral distance of 75 miles and a stratigraphic thickness of 500 feet. 21 specimen measurements from 6 samples were smeared toward the present pole. The calculations were based on the remaining 34 specimen measurements from 25 oriented samples.
	45	36 N	112 W	132½	23	7½	16	34	26 N	119 E	N	8	6½	b	a Runcorn, 1955b	
	46	36 N	112 W	132½	23	7½	16	34	24½N	120 E	N	8	4	•	b Runcorn, 1956a	
	47	36 N	113 W	133	23	8	—	—	26 N	121 E	N	9	5	c	c Irving, 1956a	
Grand Canyon, Arizona	48	—	—	146	8	7	—	—	39 N	115 E	N	—	—	b	*Doell*	*Sampling:* 42 oriented samples were collected over a lateral extent of 2 miles and a stratigraphic thickness of 500 feet. 24 of the samples showed agreement between 2 or more specimen measurements. 12 of these were smeared toward the present pole, and the calculations were based on the remaining 12 (Ref. a).
	49	36 N	112 W	146	8	7	33	12	39 N	114 E	N	7	3½	•	a Doell, 1955a b Doell, 1955b	
Hunter's Point	51	35½ N	109 W	164	5	5	32	24	50 N	96 E	N	5	3	a	*Collinson and Runcorn*	*Sampling:* 24 specimen measurements were made on 7 oriented samples.
Lower Supai, Oak Creek, Arizona	52	35 N	112 W	141	18	4	25	55	33 N	116 E	N	4	2	a	a Collinson and Runcorn, 1960	*Sampling:* 55 specimen measurements were made from 16 oriented samples.
SUPAI AVERAGE	53	—	—	—	—	9	45	7	40½N	110 E	N	9	9	•		*Sampling:* This mean pole position was found by applying Fisher statistics to the pole positions cited for entries (37), (39), (41), (46), (49), (51), and (52) above.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION E—*Continued*																
YESO FORMATION (sediments)	54	35½N	105½W	—	—	—	—	22	41 N	127 E	N	—	—	a	*Graham*	*Age:* Permian (Leonard)
	55	**35½N**	**105½W**	**143**	**−1**	**3**	**99**	**22**	**41 N**	**127 E**	**N**	**3**	**1½**	*	a Graham, 1955	*Sampling:* 26 oriented samples were collected over a lateral extent of 200 feet and a stratigraphic thickness of 36 feet. The pole position for entry (54) was scaled from Fig. 6 of Ref. a. Entry (55) was based on data scaled from Fig. 7, Ref. a; 4 widely scattered samples were excluded from the calculations.
ABO FORMATION (sediments)															*Graham*	*Age:* early Permian
Abo Canyon	56	34½N	106½W	—	—	—	—	8	44 N	120 E	N	—	—	a	a Graham, 1955	*Sampling:* 11 oriented samples were collected over a lateral distance of 400 feet and a stratigraphic thickness of 50 feet. 8 of the 20 specimen measurements made from these samples form a group and were used to give the pole position of entry (56) which was scaled from Fig. 6 of Ref. a. Entry (57) is based on data scaled from Fig. 7 of Ref. a, excluding 3 of the 20 specimen measurements.
	57	34½N	106½W	149	8	17½	5	17	42 N	117 E	N	17½	9	*		
Zuni Mts.	58	35½N	108½W	160½	55	12	7	25	17 N	87½E	N	17	12	*		*Sampling:* 13 oriented samples were collected over a lateral extent of 150 feet and a stratigraphic thickness of 25 feet. Entry (58) is based on directions of all specimens as scaled from Fig. 7, Ref. a.
ABO AVERAGE	**59**	**35 N**	**107½W**	**—**	**—**	**—**	**—**	**—**	**30 N**	**100 E**	**N**	**18**	**18**	*		*Sampling:* Average of pole positions of entries (57) and (58) is listed. 18° is the distance from average position to either pole and has no further significance.
SANGRE DE CRISTO FORMATION (sediments)	**60**	**35½N**	**105½W**	**175½**	**30½**	**11**	**9**	**19**	**38 N**	**80½E**	**N**	**11**	**6½**	*	*Graham* a Graham, 1955	*Age:* early Permian and Middle or Late Pennsylvanian *Sampling:* 19 oriented samples were collected over a lateral extent of 100 feet and a stratigraphic thickness of 10 feet. Entry (60) was based on data scaled from Fig. 7, Ref. a.
AUSTRALIA																
UPPER MARINE VOLCANIC SERIES	**61**	**35 S**	**151 E**	**67**	**81**	**11**	**—**	**—**	**27 N**	**11 W**	**N**	**21**	**21**	a	*Irving and Green* a Irving and Green, 1958 b Irving, 1959	*Age:* Cited as Permian in Ref. a *Sampling:* Ref. b states that 3 flows were sampled.
LOWER MARINE VOLCANIC SERIES	**62**	**33 S**	**151 E**	**110**	**80**	**—**	**—**	**—**	**38 N**	**6 W**	**N**	**—**	**—**	a	*Irving and Green* a Irving and Green, 1958	*Age:* Cited as Permian in Ref. a *Sampling:* Ref. b states that 1 flow was sampled.

															b Irving, 1959	
LOWER MARINE VOLCANIC SERIES	62	33 S	151 E	110	80	—	—	—	38 N	6 W	N	—	—	a	*Irving and Green* a Irving and Green, 1958 b Irving, 1959	*Age:* Cited as Permian in Ref. a *Sampling:* Ref. b states that 1 flow was sampled.
AFRICA																
MAJI YA CHUMVI FORMATION (sediments)	63	3 S	39 E	267	38	11½	9	5	4 N	150 E	N	13½	8	a	*Nairn* a Creer, Irving, Nairn, and Runcorn, 1958	*Age:* Cited as late Permian *Sampling:* 21 specimen measurements were made on 5 oriented samples.
TARU GRIT	64	3 S	39 E	87	61	16½	23	8	0 N	87 E	S	25	19½	a	*Nairn* a Creer, Irving, Nairn, and Runcorn, 1958	*Age:* Cited as early Permian *Sampling:* 32 specimen measurements were made on 8 oriented samples. *Other:* Values cited for N, α_{95}, and k are inconsistent in entries (63) and (64), and it is impossible to deduce whether specimen measurements or mean sample measurements were used in computing the statistics.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION F TRIASSIC																
EUROPE																
KEUPER MARLES															*Clegg et al.* a Clegg, Almond, and Sutbbs, 1954a b Runcorn, 1955a c Runcorn, 1955b d Runcorn, 1956a e Irving, 1956a f Day and Runcorn, 1955	*Age:* Late Triassic; sandstones sampled are close to base of Keuper Marle series
Normal sites	1	—	—	29	34	10	58	—	—	—	S	—	—	b		*Sampling:* 12, 2, 6, 3, and 2 samples, respectively, were collected at each of 5 sites over a lateral distance of 140 miles. Values from Ref. b (citing an analysis of Creer, 1955) are from p. 272 and p. 284, respectively. Calculations for entry (3) were based on mean site directions cited in Table 1, Ref. a.
	2	—	—	26	28	—	—	—	—	—	S	—	—	b		
	3	53 N	2 W	26½	34½	13½	34	5	50½N	137½E	S	15½	9	*		
																Stability: The magnetization was not changed after random orientation in the earth's field for several weeks, nor by the application of a 300-oersted A.C. field. Graham's fold test on dips of 15° and less also indicate stability.
Reversed sites	4	—	—	219	−16	27	13	—	—	—	N	—	—	b		*Sampling:* 6, 2, 2, and 8 samples, respectively, were collected at each of 4 sites over a lateral distance of 180 miles. For basis of entry (6) see remark above for entry (3).
	5	—	—	214	−28	—	—	—	—	—	N	—	—	b		
	6	53 N	2 W	218	−16	27	13	4	35½N	129½E	N	28½	15	*		
																Stability: See previous remark about stability.
Average of Normal and Reversed	7	—	—	—	—	—	—	—	48 N	155 E	M	—	—	b		*Sampling:* Values for entry (9) were scaled from Fig. 9 in Ref. d. Entry (12) was based on average directions at normal and reversed sites (Table 1, Ref. a). All entries are presumably based on these same data.
	8	—	—	—	—	—	—	—	47 N	122 E	—	13	7	c		
	9	—	—	—	—	—	—	—	46 N	131 E	—	13	7	d		
	10	53 N	2 W	33	27	12	—	—	43 N	131 E	—	12	7	e		
	11	—	—	—	—	—	—	—	47 N	133 E	—	—	—	f		
	12	**53 N**	**2 W**	**32**	**26**	**13**	**16**	**9**	**43 N**	**133 E**	**M**	**14**	**8**	*		*Other:* A detailed study of unstable Keuper Marles from another locality is given by Creer (1957a).
VOSGES SANDSTONE	**13**	**48 N**	**6 E**	**218**	**16**	**12**	**2**	**61**	**27½N**	**142 E**	**N**	**12**	**6**	*	*Clegg et al.* a Clegg, Deutsch, Everitt, and Stubbs, 1957 b Irving, 1959	*Age:* Cited as Early Triassic in Ref. a
	14	—	—	—	—	—	—	—	28 N	143 E	N	12	6	b		*Sampling:* 61 specimens or samples (?) were collected at 7 sites. Entry (13) was based on data scaled from Fig. 5, Ref. a.
VILLAVICIOSA SANDSTONE	**15**	**43½N**	**5½W**	**4**	**56**	**2**	**80**	**87**	**82 N**	**150 E**	**S**	**2½**	**2**	*	*Clegg et al.* a Clegg, Deutsch, Everitt, and Stubbs, 1957	*Age:* Cited as Triassic in Ref. a *Sampling:* It is not known whether 87 is the number of oriented samples or specimen measurements. Entry (15) was based on data scaled from Fig. 6, Ref. a.
																Stability: Stability is suggested by the tightness of the group which *in situ* (*i.e.*, before correction for geologic dip) has a direction 30° from the present field

SANDSTONE														Everitt, and Stubbs, 1957	of oriented samples or specimen measurements. Entry (15) was based on data scaled from Fig. 6, Ref. a.
															Stability: Stability is suggested by the tightness of the group which *in situ* (*i.e.*, before correction for geologic dip) has a direction 30° from the present field direction. Of 7 sites sampled in this region, only Villaviciosa, Alcolea, and Aguilar had consistency of directions of magnetization.
ALCOLEA AND AGUILAR SANDSTONE	**16**	**41 N**	**2½W**	**349½**	**51½**	**5½**	**50**	**16**	**78 N**	**135 W**	**S**	**7½**	**5½**	•	*Clegg et al.* a Clegg, Deutsch, Everitt, and Stubbs, 1957	*Age:* Cited as Triassic in Ref. a *Sampling:* 16 samples or specimens (?) were collected at 2 sites. Entry (16) was based on data scaled from Fig. 7, Ref. a.
VETLUJSKIJ SEDIMENTS	17	59 N	50 E	235	21	—	—	9	—	—	—	—	—	a	*Khramov* a Khramov, 1958	*Age:* Early Triassic *Sampling:* 9 oriented samples were collected over a lateral extent of 100 km and a thickness of 40 m. Entries (19) and (20) refer to the direction of the "stable" (Ref. a) component. Much scatter is present in the plotted data.
	18	59 N	50 E	235	21	—	—	9	7½N	176 E	N	—	—	•		
	19	59 N	50 E	222	−19	—	—	—	—	—	—	—	—	a		
	20	**59 N**	**50 E**	**222**	**−19**	**—**	**—**	**—**	**31 N**	**179½E**	**N**	**—**	**—**	•		
NORTH AMERICA																
SPRINGDALE SANDSTONE MEMBER OF MOENAVE FORMATION	21	37½N	113 W	350	39	8½	17	18	—	—	S	—	—	a	*Runcorn* a Runcorn, 1956a b Runcorn, 1955b	*Age:* Late Triassic (?) *Sampling:* 8 oriented samples were collected at 1 locality. The statistical analysis of entry (21) is based on 18 specimens from 7 of these samples. The pole position cited in entry (24) is suggested (Ref. a) as more probable owing to "smearing" toward the present pole.
	22	**37½N**	**113 W**	**350**	**39**	**8½**	**17**	**18**	**72½**	**98 E**	**S**	**10**	**6**	•		
	23	—	—	—	—	—	—	—	60 N	110 E	S	—	—	b		
	24	—	—	338	16	—	—	—	55 N	107 E	S	—	—	a		
REDONDA FORMATION (sediments)	25	35 N	104 W	16½	55½	4½	57	20	77 N	23 W	S	6½	4½	•	*Graham* a Graham, 1955	*Age:* Late Triassic, equivalent to the upper part of the Chinle *Sampling:* 20 specimen measurements were made on 17 oriented samples collected over a lateral extent of ½ mile and a stratigraphic thickness of 150 feet. Calculations for entry (25) were based on data scaled from Fig. 7, Ref. a.
CHINLE FORMATION (sediments) Romeroville, New Mexico	26	35½N	105 W	15½	9	9	14	16	56 N	47 E	M	9	4½	•	*Graham* a Graham, 1955	*Age:* Late Triassic. The Chinle includes the Redonda formation (upper portion) and the Shinarump member of the Chinle formation (lower portion). *Sampling:* 16 oriented samples collected at 1 site over a horizontal distance of 100 feet and a vertical thickness of 15 feet. Entry (26) was based on data scaled from Fig. 7, Ref. a. *Reversals:* 1 specimen is reversely magnetized.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION F—*Continued*																
Las Vegas, New Mexico	27	35½ N	105 W	48	32	—	—	—	—	—	—	—	—	b	*Graham*	*Sampling:* 8 specimen measurements were made on 6 oriented samples. Lateral sampling extent is 100 feet, vertical sampling extent is 5 feet. Entry (28) is based on data scaled from Fig. 7 of Ref. a.
	28	35½	105 W	33	47½	16½	12	8	61 N	10 W	S	21½	14	*	a Graham, 1955 b Kintzinger, 1957	
Site 1, Moab, Utah	29	38½ N	109½ W	160	44	—	—	39	23 N	90 E	N	—	—	a	*Collinson and Runcorn*	*Sampling:* 39 specimen measurements were made on 14 oriented samples.
	30	38½ N	109½ W	156	−8	—	—	39	49 N	109 E	N	—	—	a	a Collinson and Runcorn, 1960	*Stability:* Because of "smearing" these samples are considered to be only partially stable; entry (30) is an estimate (Ref. a) of the stable component of magnetization.
Site 2, Moab, Utah	31	38½ N	109½ W	59	73	—	—	60	48 N	66 W	S	—	—	a	*Collinson and Runcorn*	*Sampling:* 60 specimen measurements were made on 10 oriented samples.
	32	38½ N	109½ W	160	10	—	—	60	50 N	114 E	N	—	—	a	a Collinson and Runcorn, 1960	*Stability:* Because of "smearing" these samples are considered to be only partially stable; entry (32) is an estimate (Ref. a) of the stable component of magnetization.
Site 1, Colo. National Monument	33	39 N	108½ W	356	66	5	25	29	—	—	—	—	—	a	*Collinson and Runcorn*	*Sampling:* 29 specimen measurements were made on 6 oriented samples.
	34	39 N	109 W	356	66	5	25	29	80½N	125 W	S	8½	7	*	a Collinson and Runcorn, 1960	*Stability:* Cited as unstable (Ref. a)
Site 2, Colo. National Monument	35	39 N	108½ W	34	60	7	14	31	—	—	—	—	—	a	*Collinson and Runcorn*	*Sampling:* 31 specimen measurements were made on 7 oriented samples.
	36	39 N	108½ W	34	60	7	14	31	64 N	35 W	S	10½	8	*	a Collinson and Runcorn, 1960	*Stability:* Cited as unstable (Ref. a)
SHINARUMP MEMBER OF CHINLE FORMATION (sediments)	37	36 N	111½W	355	43	6½	27	17	78½N	90 E	M	8	5	*	*Graham*	*Age:* Late Triassic
	38	36 N	111½W	356	33	—	—	17	—	—	—	—	—	b	a Graham, 1955 b Kintzinger, 1957	*Sampling:* 20 specimen measurements were made on 17 oriented samples taken over a lateral extent of 20 feet and a stratigraphic thickness of 5 feet. Entry (37) was based on data scaled from Fig. 7, Ref. a. 3 widely scattered specimen measurements were excluded.
																Reversals: 4 specimens are reversely magnetized.
CHINLE AVERAGE	39	—	—	—	—	27½	5	8	78 N	16 E	M	27½	27½	*		*Sampling:* These values were calculated by applying Fisher statistics to the pole positions cited in entries (25), (26), (28), (30), (32), (34), (36), and (37). These

CHINLE AVERAGE	39	—	—	—	—	27½	5	8	78 N	16 E	M	27½	27½	*		*Sampling:* These values were calculated by applying Fisher statistics to the pole positions cited in entries (25), (26), (28), (30), (32), (34), (36), and (37). These include the poles for the equivalent Redonda formation and the Shinarump member of the Chinle formation. *Other:* Owing to "smearing" and instability at some sites (see above) the mean pole position cited here is probably too far north. The confidence intervals for the data used in this computation vary widely, and the value of α_{95} listed is therefore of doubtful significance.
CHUGWATER FORMATION (sediments)															*Collinson and Runcorn*	*Age:* Cited as Triassic in Ref. a; generally taken as Triassic and Permian
Troublesome Creek, Wyoming	41	42 N	106½ W	134	−12	7	10	43	36 N	135 E	N	7	4	a	a Collinson and Runcorn, 1960	*Sampling:* 43 specimen measurements were made on 10 oriented samples.
Alcova Reservoir, Wyoming	42	42½ N	107 W	152	−18	8	10	48	49 N	118 E	N	8	4	a		*Sampling:* 48 specimen measurements were made on 10 oriented samples.
Sheep Mtn., Wyoming	43	44½ N	108 W	148	−6	3	58	35	40 N	115 E	N	3	2	a		*Sampling:* 35 specimen measurements were made on 9 oriented samples.
Shell, Wyoming	44	44½ N	107½ W	155	−23	7	16	16	—	—	N	—	—	a		*Sampling:* 5 specimen measurements for the normal group were made on 1 oriented sample. 16 specimen measurements for the reversed group were made on 4 oriented samples. Entry (48) is an average for the normal and reversed groups.
	45	44½ N	107½ W	155	−23	7	16	16	50½N	113 E	N	8	4	*		
	46	44½ N	107½ W	340	36	—	—	5	—	—	S	—	—	a		
	47	44½ N	107½ W	340	36	—	—	5	60 N	110½E	S	—	—	*		
	48	44½ N	107½ W	—	—	—	—	—	56 N	113 E	M	8	4	a		
Fort Washakie, Wyoming	49	43 N	109 W	146	6	5	34	28	—	—	N	—	—	a		*Sampling:* 25 specimen measurements for the normal group were made on 6 oriented samples. 28 specimen measurements for the reversed group were made on 5 oriented samples. Entry (53) is an average for the normal and reversed groups.
	50	43 N	109 W	146	6	5	34	28	34½N	114 E	N	4½	2½	*		
	51	43 N	109 W	346	14	6	22	25	—	—	S	—	—	a		
	52	43 N	109 W	346	14	6	22	25	51½N	94½E	S	6½	3½	*		
	53	43 N	109 W	—	—	—	—	—	44 N	105 E	M	5	3	a		
Dinwoody Lake, Wyoming	54	43½ N	109½ W	154	−20	7	17	22	—	—	N	—	—	a		*Sampling:* The 22 specimen measurements for the reversed group were made on 6 oriented samples. The 20 specimen measurements for the normal group were made on 4 oriented samples. Entry (58) is an average for the normal and reversed groups.
	55	43½ N	109½ W	154	−20	7	17	22	50 N	112 E	N	8	4	*		
	56	43½ N	109½ W	339	25	6	30	20	—	—	S	—	—	a		
	57	43½ N	109½ W	339	25	6	30	20	55 N	107½E	S	6½	3½	*		
	58	43½ N	109½ W	—	—	—	—	—	52 N	110 E	M	5	3	a		
Lander, Wyoming	59	42½ N	108½ W	157	15	9	8	37	—	—	N	—	—	a		*Sampling:* The 37 specimen measurements for the reversed group were made on 8 oriented samples. The 12 specimen measurements for the normal group were made on 2 oriented samples. Entry (63) is an average for the normal and reversed groups.
	60	42½ N	108½ W	157	15	9	8	37	35 N	99 E	N	9½	4½	*		
	61	42½ N	108½ W	344	27	4	55	12	—	—	S	—	—	a		
	62	42½ N	108½ W	344	27	4	55	12	58 N	101 E	S	5	2½	*		
	63	42½ N	108½ W	—	—	—	—	—	47 N	100 E	M	8	4	a		

Reversals: 4 specimens are reversely magnetized.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION F—*Continued*																
Thermopolis, Wyoming	64	43½ N	108 W	170	−19	11	13	15	—	—	N	—	—	a		*Sampling:* 18 specimen measurements for the normal group were made on 3 oriented samples. 15 specimen measurements for the reversal group were made on 3 oriented samples. Entry (68) is an average for the normal and reversed groups.
	65	43½ N	108 W	170	−19	11	13	15	54½N	90 E	N	11½	6	*		
	66	43½ N	108 W	332	16	5	56	18	—	—	S	—	—	a		
	67	43½ N	108 W	332	16	5	56	18	46½N	114 E	S	5½	3	*		
	68	43½ N	108 W	—	—	—	—	—	52 N	103 E	M	7	4	a		
Rawlins, Wyoming	69	42 N	107½ W	149	−19	6	25	21	—	—	N	—	—	a		*Sampling:* 21 specimen measurements for the reversed group were made on 6 oriented samples. 12 specimen measurements for the normal group were made on 3 oriented samples. Entry (73) is an average for the normal and reversed groups.
	70	42 N	107½ W	149	−19	6	25	21	47½N	122 E	N	7	3½	*		
	71	42 N	107½ W	328	12	4	43	12	—	—	S	—	—	a		
	72	42 N	107½ W	328	12	4	43	12	44 N	120½E	S	3½	2	*		
	73	42 N	107½ W	—	—	—	—	—	46 N	121 E	M	5	3	a		
Red Mtn., Wyoming	74	41 N	106 W	151	−6	6	12	52	—	—	N	—	—	a		*Sampling:* 52 specimen measurements for the reversed group were made on 13 oriented samples. 16 specimen measurements for the normal group were made on 3 oriented samples. 2 samples were not used in the calculations. Entry (78) is an average for the normal and reversed groups.
	75	41 N	106 W	151	−6	6	12	52	44 N	116 E	N	6	3	*		
	76	41 N	106 W	335	15	6	30	16	—	—	S	—	—	a		
	77	41 N	106 W	335	15	6	30	16	50 N	114 E	S	6½	3	*		
	78	41 N	106 W	—	—	—	—	—	47 N	116 E	M	5	3	a		
CHUGWATER AVERAGE	**79**	—	—	—	—	**5**	**57**	**17**	**48 N**	**112½E**	**M**	**5**	**5**	*		*Sampling:* These calculations were made by applying Fisher statistics to the pole positions given for each site; reversed groups are considered as separate sites. The 17 entries used are (41), (42), (43), (45), (47), (50), (52), (55), (57), (60), (62), (65), (67), (70), (72), (75), and (77).
MOENKOPI FORMATION (Sediments)																*Age:* Early to Middle (?) Triassic
Zion Nat. Park	80	36 N	111½W	0	27½	—	—	—	—	—	—	—	—	a	*Runcorn* a Runcorn, 1956a	*Sampling:* 18 oriented samples were collected over a lateral extent of 250 miles. Entry (81) was based on mean directions of 16 samples (Table 6, Ref. a); two samples with widely divergent directions were not included.
	81	36 N	111½W	1½	28	13	9	16	69 N	64 E	S	14½	8	*		
Marble Canyon	82	37 N	111½W	325	35	—	—	11	—	—	—	—	—	a	*Kintzinger* a Kintzinger, 1957	*Sampling:* 11 oriented samples were collected over a lateral extent of 1 mile and a stratigraphic thickness of 230 feet. The values cited for entries (82) and (83) appear in a table and were scaled from
	83	37 N	111½W	340	28	—	—	11	—	—	—	—	—	a		
	84	37 N	111½W	345	29½	15½	11	11	65 N	105 E	S	17	9½	*		

Marble Canyon	82	37 N	111½W	325	35	—	—	11	—	—	—	—	—	a	*Kintzinger* a Kintzinger, 1957	*Sampling:* 11 oriented samples were collected over a lateral extent of 1 mile and a stratigraphic thickness of 230 feet. The values cited for entries (82) and (83) appear in a table and were scaled from Fig. 1, respectively, in Ref. a. Entry (84) was based on data scaled from Fig. 1 in Ref. a.
	83	37 N	111½W	340	28	—	—	11	—	—	—	—	—	a		
	84	37 N	111½W	345	29½	15½	11	11	65 N	105 E	S	17	9½	*		
Echo Cliffs, Arizona	86	37 N	111½ W	349	28	6	17	42	66 N	95 E	S	6	4	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Sampling:* The statistical analysis was based on 42 specimen measurements made on 18 oriented samples; 6 specimen measurements on 4 samples were discarded.
Poverty Tank, Arizona	87	36 N	111½ W	337	36	7	22	27	64 N	127 E	S	8	5	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Sampling:* The statistical analysis was based on 27 specimen measurements made on 10 oriented samples; 3 specimen measurements from 1 sample discarded.
Vernal, Utah	88	40½ N	109½ W	158	−4	6	18	38	46 N	103 E	N	6	3	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Sampling:* The statistical analysis was based on 38 specimen measurements made on 8 oriented samples. 3 specimen measurements from 1 sample were discarded.
Split Mtn., Colorado	89	40½ N	109 W	156	−4	6	28	23	46 N	106 E	N	6	3	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Sampling:* 23 specimen measurements were made on 5 oriented samples.
Sand Canyon, Colorado	90	40½ N	109 W	148	−7	9	10	27	43 N	118 E	N	9	4	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Sampling:* The statistical analysis was based on 27 specimen measurements made on 7 oriented samples. 3 specimen measurements from 1 sample were discarded.
Moenkopi Average	**91**	—	—	—	—	**12½**	**20**	**8**	**62 N**	**103 E**	**M**	**12½**	**12½**	*		*Sampling:* These calculations were made by applying Fisher statistics to the pole positions of entries (81), (84), (86), (87), (88), (89), and (90). Data on partially unstable Capitol Reef outcrop (Ref. a) not included.
Connecticut Valley Massachusetts Lavas	92	42 N	73 W	10	14	11	—	8	54 N	90 E	S	8	6	a	*Du Bois et al.* a Du Bois, Irving, Opdyke, Runcorn, and Banks, 1957	*Age:* Cited as Triassic *Sampling:* 8 oriented samples were collected from 3 flows. Entry (93) is a minimum value of α_{95} based on 3 points in time (3 flows) and a "secular variation precision" $k' = 30$.
	93	**42 N**	**72½W**	**10**	**14**	**15**	**k'**	**3**	**54 N**	**90 E**	**S**	**15**	**7½**	*		
Connecticut Lavas and Sediments	94	42 N	73 W	12	14	15	—	12	55 N	88 E	—	15	8	a	*Du Bois et al.* a Du Bois, Irving Opbyke, Runcorn, and Banks, 1957 d Du Bois, 1957	*Age:* Cited as Triassic in Ref. a *Sampling:* 12 oriented samples were collected (Ref. a); 32 specimens were measured (Ref. b). The value of k for entry (96) was calculated by the approximate formula. *Reversals:* A plot in Ref. b shows that about half the specimens are reversed.
	95	—	—	12	14	3	—	32	41 N	91 E	M	—	—	b		
	96	**42 N**	**73 W**	**12**	**14**	**15**	**7**	**12**	**53 N**	**86 E**	**M**	**15**	**7½**	*		

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION F—*Continued*																
BRUNSWICKIAN FORMATION* (sediments)	97	41 N	75 W	6	28	3	—	71	63 N	93 E	S	6	3	a	*Du Bois et al.*	*Age:* Late Triassic, top of the Newark group; referred to as "Newark formation" in Ref. b.
	98	**40½N**	**75 W**	**2**	**22**	**3**	**86**	**70**	**61 N**	**102 E**	**S**	**3**	**2**	**b**	a Du Bois, Irving, Opdyke, Runcorn, and Banks, 1957 b Collinson and Runcorn, 1960	*Sampling:* 71 specimen measurements were made on 21 oriented samples (Ref. a).
NEW OXFORD FORMATION (sediments)	99	39½N	77½W	—	—	—	—	—	63 N	158 E	S	—	—	a	*Graham*	*Age:* Late (?) Triassic, bottom of the Newark group
	100	**39½N**	**77½W**	**334**	**48**	**6½**	**36**	**12**	**66 N**	**174 E**	**S**	**8½**	**5½**	*	a Graham, 1955	*Sampling:* 13 oriented samples were collected over a lateral distance of 300 feet and a stratigraphic thickness of 20 feet. Values for entry (100) were based on data scaled from Fig. 7, Ref. a. One sample is not included. The pole position for entry (99) was scaled from Fig. 6, Ref. a.
NOVA SCOTIA LAVAS	101	—	—	—	—	5½	—	74	77½N	72 E	—	5½	5½	a	*Bowker*	*Age:* Cited as Triassic in Ref. a
	102	**—**	**—**	**—**	**—**	**11½**	**14**	**12**	**77 N**	**75½E**	**—**	**11½**	**11½**	*	a Bowker, 1959 (letter of Nov. 16, 1959 to Doell)	*Sampling:* 217 specimen measurements were made on 74 oriented samples collected at 12 sites. The values for entry (102) were based on averages for each site. In both entries, the statistical calculations were made on pole positions rather than magnetic directions. *Stability:* A.C. partial demagnetization decreased the scatter.
DINOSAUR CANYON SANDSTONE MEMBER OF THE MOENAVE FORMATION	103	37 N	111½W	21	30	—	—	—	—	—	—	—	—	a	*Kintzinger*	*Age:* Late Triassic (?); younger than the Chinle formation
	104	37 N	111½W	13½	27	—	—	—	—	—	—	—	—	a	a Kintzinger, 1957	*Sampling:* Sampling extent was over 1 mile laterally and 80 feet stratigraphically. Values for entry (103) are from the table in Ref. a, and entry (104) was scaled from the plot in the same Ref. Entry (105) is based on individual sample data scaled from Fig. 1 in Ref. a.
	105	**37 N**	**111½W**	**17**	**31**	**11**	**12**	**17**	**65 N**	**28 E**	**S**	**12**	**6½**	*		

* *Editor's Note:* Data for No. 98 withdrawn from Ref. b in final version of manuscript.

AUSTRALIA																
TASMANIAN VOL-	**106**	**42 S**	**147 E**	**—**	**−81½**	**—**	**—**	**17**	**42 N**	**33 W**	**M**	**>16**	**>16**	*	*Almond et al.*	*Age:* The tuffs underlie the Tasmanian dolerite and

Fig. 1 in Ref. a.

* *Editor's Note:* Data for No. 98 withdrawn from Ref. b in final version of manuscript.

AUSTRALIA															
TASMANIAN VOLCANIC TUFFS	106	42 S	147 E	—	−81½	—	—	17	42 N	33 W	M	>16	>16	*	*Almond et al.* a Almond, Clegg, and Jaeger, 1956
BRISBANE TUFF	107	27½S	153½E	35	−83	—	—	—	39 N	37 W	M	—	—	a	*Irving and Green* a Irving and Green, 1958
AFRICA															
BECHUANALAND CAVE SANDSTONE	108	—	—	326	−16	9½	6	44	54 N	79 W	S	—	—	a	*Nairn* a Nairn, 1957a b Creer, Irving, Nairn and Runcorn, 1958
	109	23 S	27 E	326	−16	9½	6	44	54 N	44 W	S	10	5	b	

Age: The tuffs underlie the Tasmanian dolerite and are Triassic(?)
Sampling: 20 feet of azimuthally unoriented core was sampled. The inclination indicated is an average of data in Table 4, Ref. a.
Other: Since $I = 81\frac{1}{2}°$ corresponds to latitude 74°, the pole should be roughly 16° away from the sampling area.
Reversals: 2 of the 17 samples have reversed polarity.

Age: Cited as probably Early Triassic
Reversals: Reversals are noted, but no details are given.

Age: Late Triassic, immediately underlying the Karroo basalts (Du Toit, 1953, p. 301)
Sampling: 44 specimen measurements were made on 8 oriented samples.
Other: Contemporaneous sediments from adjacent regions have scattered directions.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION G JURASSIC																
EUROPE																
NORTH YORKSHIRE SEDIMENTS	1	—	—	3½	67	2½	88	36	—	—	—	—	—	a	*Nairn*	*Age:* Early Jurassic (middle Lias–lower Corallian)
	2	54 N	1 W	3½	67	4½	88	36	85 N	150 E	S	7½	6	*	a Nairn, 1956 b Nairn, 1957c	*Sampling:* 36 specimen measurements were made on 6 oriented samples. These were later rejected as unstable (Ref. b), because α_{95} changed from 4.7 to 6.7 after 18 months storage. The mean was little affected. *Stability:* Since the circle of confidence includes the present pole there is no evidence for stability.
SEDIMENTS OF SCOTLAND	3	—	—	234½	−66	6½	33	14	56 N	76 E	N	—	—	a	*Nairn*	*Age:* Cited as Early Jurassic
	4	—	—	226	−65	6½	33	14	—	—	—	—	—	b	a Nairn, 1957c	*Sampling:* Entry (3) was based on 14 specimen measurements made on 4 oriented samples, 3 from the limestone of the Estuarine beds and 1 from the Brora Arenaceous series; these samples had reversed magnetizations. Entry (5) was based on average sample directions of the 4 samples plus an additional 2 samples from the Lias limestones, one of which is normal and the other reversed (Table III, Ref. a).
	5	57½N	5 W	220½	−71	18	14	6	68 N	73 E	M	31½	27	*	b Nairn, 1957a	
MIDFORD SANDS	6	51½N	2½W	103½	70	7½	10	42	33 N	41 E	S	13	11	a	*Girdler* a Girdler, 1960	*Age:* Early Jurassic (upper Lias) *Sampling:* 42 specimen measurements were made on 20 oriented samples collected at 3 sites separated by more than 1 mile. *Stability:* Scatter was reduced by application of Graham's fold test. Specimens were remeasured after 6 months with no change in magnetization. *Other:* Samples at nearby sites were unstable.
COTSWOLD SEDIMENTS	7	51½N	2½W	262½	−64	10	7	38	38 N	59½ E	N	15½	12½	a	*Girdler* a Girdler, 1960	*Age:* Early Jurassic (upper Lias); older than the Midford sands *Sampling:* 38 specimen measurements were made on 17 oriented samples collected at 2 sites. *Stability:* Specimens were remeasured after 6 months with no change in magnetization.
PYRENEES VOLCANIC ROCKS	8	43 N	1½E	55	60	6	17	38	49 N	76½E	S	8½	6½	a	*Girdler* a Girdler, 1960	*Age:* Early Jurassic (lower Lias), on the basis of intercolated limestone beds *Sampling:* 38 specimen measurements were made on

PYRENEES VOLCANIC ROCKS	8	43 N	1½E	55	60	6	17	38	49 N	76½E	S	8½	6½	a	*Girdler* a Girdler, 1960	*Age:* Early Jurassic (lower Lias), on the basis of intercolated limestone beds *Sampling:* 38 specimen measurements were made on 18 oriented samples collected from 4 sites separated by more than 1 mile. *Stability:* Correction for rather uniform dips up to 40° did not significantly reduce the scatter. *Other:* The value of a_{95} is probably too high because individual specimen measurements were used.
ALPINE RADIOLARITE	9	47½N	12½E	36½	48	5½	100	21	58½N	128 E	S	—	—	a	*Hargraves and Fischer* a Hargraves and Fischer, 1959	*Age:* Cited as Middle Jurassic *Sampling:* 21 specimen measurements were made on 15 oriented samples taken at one site over a stratigraphic thickness of 7 meters.
	10	47½N	12½E	36½	48	5½	100	21	56 N	122½E	S	7	4½	*		
ALPINE LIMESTONE	11	47½N	12½E	48	50½	6½	71	30	53 N	112 E	S	—	—	a	*Hargraves and Fischer* a Hargraves and Fischer, 1959	*Age:* Cited as Early Jurassic *Sampling:* 30 specimen measurements were made on 16 oriented samples taken at 1 site over a stratigraphic thickness of 4 m.
	12	47½N	12½E	48	50½	6½	71	30	50 N	109½E	S	9	6	*		
NORTH AMERICA																
KAYENTA FORMATION																*Age:* The Kayenta is Early Jurassic (?) (Glen Canyon group).
Central Arizona	13	—	—	353	43½	9½	50	6	—	—	S	—	—	a	*Runcorn* a Nairn, 1956	
	14	35 N	111 W	353	43½	9½	50	6	79 N	102 E	S	12	7½	*		
Kayenta Arizona	15	36½ N	110½ W	20	50	6	14	43	72 N	6 W	S	8	6	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Sampling:* 43 specimen measurements were made on 8 oriented samples.
Echo Cliffs, Arizona	16	37 N	111½ W	14	43	6	14	43	73 N	19 E	S	7	5	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Sampling:* 43 specimen measurements were made on 14 oriented samples.
Navajo National Monument	17	36½ N	111 W	335	50	11	7	27	—	—	S	—	—	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Sampling:* 27 specimen measurements were made on 7 oriented samples.
	18	36½ N	111 W	335	50	11	7	27	68½ N	151 E	S	15	10	*		
Kanab, Utah	19	37 N	112½ W	4	53	9	12	33	85 N	23 E	S	13	9	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Sampling:* The statistical analysis was based on 33 specimen measurements on 8 oriented samples. 6 specimen measurements from 2 oriented samples were discarded. The circle of confidence includes the present dipole direction.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION G—*Continued*																
KAYENTA AVERAGE	20	—	—	—	—	21	20	4	83 N	63 E	S	21	21	*		*Sampling:* These values were computed by applying Fisher statistics to the pole positions for entries (14), (15), (16), (18), and (19).
CARMEL FORMATION (sediments)	21	38½ N	109½ W	349	63	9	10	31	80 N	160 W	S	14	11	a	*Collinson and Runcorn* a Collinson and Runcorn, 1960	*Age:* Cited as Jurassic in Ref. a *Sampling:* 31 specimen measurements were made on 9 oriented samples.
AFRICA																
KARROO DOLERITES															*Graham and Hales* a Graham and Hales, 1957	*Age:* The dolerites were emplaced during and somewhat after extrusion of the Karroo basalts (Du Toit, 1953, p. 369). Age is Late Triassic (Walker and Poldervaart, 1949, p. 598–599) or Early Jurassic (Du Toit, 1953, p. 370).
Winkelhaak Upper sill	22	26½S	29 E	173	58	5	—	17	77 N	127 W	N	7	6	a		*Sampling:* 17 oriented samples were taken in mine shafts. A 75-foot thickness of the sill was sampled at 4 sites a maximum of 1 mile apart.
Winkelhaak Lower sill	23	26½S	29 E	307	−63	12	—	8	44 N	98 W	S	18	14	a		*Sampling:* 8 oriented samples were taken in mine shafts. A 50-foot section of the sill was sampled at 4 sites a maximum of 1 mile apart.
Estcourt reversed dolerite	24	29 S	30 E	167	51	6	—	15	77 N	90 W	N	8	6	a		*Sampling:* 15 oriented samples were collected in a tunnel about 1 mile long.
	25	29 S	30 E	167	51	6	—	15	79 N	79 W	N	8	6	*		
Estcourt normal dolerite	26	29 S	30 E	331	−64	8	—	9	67 N	86 W	S	12	10	a		*Sampling:* 9 oriented samples were collected in a tunnel about 1 mile long.
	27	29 S	30 E	331	−64	8	—	9	62 N	105 W	S	13	10	*		
DOLERITE AVERAGE	28	—	—	—	—	20	23	4	66 N	101½W	M	20	20	*		*Sampling:* These values were computed by applying Fisher statistics to the pole positions of entries (22), (23), (25), and (27).
ESTCOURT BAKED SEDIMENTS																
Reversed	29	29 S	30 E	180	55	11	—	8	—	—	—	—	—	a	*Graham and Hales* a Graham and Hales, 1957	*Sampling:* These samples were baked by the reversed and normal dolerites at Estcourt, respectively, and were collected in the same tunnel. They are in good

Average	35															Fisher statistics to the pole positions of entries (22), (23), (25), and (27).
Estcourt Baked Sediments															*Graham and Hales*	*Sampling:* These samples were baked by the reversed and normal dolerites at Estcourt, respectively, and were collected in the same tunnel. They are in good agreement with the dolerite samples in both cases.
Reversed	29	29 S	30 E	180	55	11	—	8	—	—	—	—	—	a	a Graham and Hales, 1957	
Normal	30	29 S	30 E	324	−62	7	—	7	—	—	—	—	—	a		
Karroo Surface Samples	**31**	**30½S**	**28½E**	**352**	**−62**	**12**	**—**	**33**	**76 N**	**128 W**	**M**	**19**	**14**	**a**	*Graham and Hales*	*Sampling:* 33 oriented samples were collected over an area 250 by 400 miles.
															a Graham and Hales, 1957	*Reversals:* 8 of the 33 samples had reversed polarity.
Karroo Basalts	32	—	—	330	−40½	4½	19	44	61 N	76 W	S	—	—	a	*Nairn*	*Age:* The Karroo basalts are the top of the Stormberg series immediately overlying the Cave sandstone. The age is Late Triassic (Walker and Poldervaart, 1949, p. 598–599) or Early Jurassic (Du Toit, 1953, p. 370).
	33	—	—	332	−40	4½	19	44	—	—	—	—	—	b	a Nairn, 1956	
	34	18 S	26 E	332	−40	5	19	44	63 N	78 W	S	6	2½	c	b Nairn, 1957a	
	35	18 S	26 E	328	−40	4½	19	44	60 N	77 W	S	5½	3	d	c Creer, 1958	
	36	**18 S**	**26 E**	**332**	**−40**	**9½**	***k′***	**7**	**63½ N**	**79 W**	**S**	**11½**	**7**	•	d Creer, Irving, Nairn, and Runcorn, 1958	*Sampling:* 44 specimen measurements were made on 11 oriented samples collected from 6 or 7 flows. Values for entry (36) are based on the "secular variation precision" $k' = 30$ and $N = 7$, the number of flows.
																Other: The pole position at latitude 2 N, longitude 8 W mentioned in Ref. b apparently is a misprint.
SOUTH AMERICA																
Parana Basin Basalts															*Creer*	*Age:* Exact age within the Mesozoic is not known
															a Creer, 1958	
Normal group	37	29 S	57 W	348	−47	2	233	22	—	—	—	—	—	a		*Sampling:* The 48 specimen measurements (normal and reversed) were made on 12 oriented samples collected at 4 sites.
	38	29 S	57 W	348	−47	2	233	22	80 N	146 W	S	2½	1½	•		
Reversed group	39	29 S	57 W	174	39	7	18	26	—	—	—	—	—	a		*Stability:* The scatter in directions was considerably decreased by partial demagnetization in A.C. fields of 250 oersted.
	40	29 S	57 W	174	39	7	18	26	81 N	96 W	N	8½	5	•		
Basalt average	41	29 S	57 W	351	−42	2	82	48	—	—	—	—	—	a		
	42	29 S	57 W	351	−42	2	82	48	81 N	118 W	M	2½	1½	•		
Tacuarembo Baked Sandstone															*Creer*	*Age:* These sandstones were baked by the Parana Basin basalts. Thus the age of their magnetization within the Mesozoic is not known.
															a Creer, 1958	
Normal group	43	29 S	57 W	357	−42	3	25	71	—	—	—	—	—	a		*Sampling:* The 81 specimen measurements (normal and reversed) were made on 12 oriented samples collected at 4 sites.
	44	29 S	57 W	357	−42	3	25	71	84 N	96 W	S	4	2	•		
Reversed group	45	29 S	57 W	176	43	14½	12	10	—	—	—	—	—	a		
	46	29 S	57 W	176	43	14½	12	10	77 N	132 W	N	18	11	•		
Sandstone	47	29 S	57 W	356	−43	3½	22	81	—	—	—	—	—	a		
Average	48	29 S	57 W	356	−43	3½	22	81	84½ N	102 W	M	4	2½	•		

UMI

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION G—*Continued*																
PARANA BASALT AND SANDSTONE	49	29 S	57 W	354	−43	2½	24	129	83 N	126 W	M	3	1½	a	*Creer*	*Sampling:* For entry (49) Fisher statistics were applied to the 129 specimen measurements (basalts and sandstones) made on the 24 oriented samples (Ref. a). Entries (50), (51), and (52) are based on the "secular variation precision" $k' = 30$ and the 3 following assumptions about the number of different points in time that were sampled. Entry (50): the 12 sandstone samples were baked by 12 different flows, and in addition 12 other lava flows were also sampled. Entry (51): the sandstone samples were baked by 12 different flows, and in all cases both a flow and its baked sediment were sampled. Entry (52): only 1 lava flow and its associated baked sediment were sampled at each of the 4 sites. Ref. a does not specify the number of flows sampled.
	50	29 S	57 W	354	−43	5	k'	24	83½ N	112 W	M	6½	4	•	a Creer, 1958	
	51	29 S	57 W	354	−43	7½	k'	12	83½ N	112 W	M	9	5½	•		
	52	**29 S**	**57 W**	**354**	**−43**	**12½**	k'	**4**	**83½ N**	**112 W**	**M**	**16**	**9½**	•		
SECTION H CRETACEOUS																
EUROPE																
WEALDEN SEDIMENTS	1	—	—	345	63	—	—	—	—	—	—	—	—	a	*Wilson*	*Age:* Early Cretaceous *Sampling:* 97 specimen measurements on 19 oriented samples collected over a lateral extent of 4 miles were made. Mean sample directions were used in the statistical analysis, and the data for entry (2) were calculated from other statistical parameters reported in Ref. a. *Stability:* Correction for small-amplitude folding reduced scatter. Partial A.C. demagnetization at 180 oersted reduced scatter. Sands and clays of different lithology and different intensities had closely parallel directions. *Other:* 27 additional cretaceous samples were rejected as unstable.
	2	**50½N**	**1½W**	**345**	**63**	**2**	**260**	**19**	**79 N**	**117 W**	**S**	**3**	**2**	•	a Wilson, 1959	
NORTH AMERICA																
DAKOTA SAND-	3	34 N	110 W	164	−62	—	—	10	76½	127 E	N	11	8½	a	*Runcorn*	*Age:* Lies above the Trinity group; Early(?) and Late

Other: 27 additional cretaceous samples were rejected as unstable.																
NORTH AMERICA																
DAKOTA SANDSTONE	3	34 N	110 W	164	−62	—	—	10	76½	127 E	N	11	8½	a	*Runcorn*	*Age:* Lies above the Trinity group; Early(?) and Late Cretaceous
	4	**34 N**	**110 W**	**169**	**−57**	**15**	**17**	**5**	**80 N**	**176 W**	**M**	**21½**	**16**	*	a Runcorn, 1956a	*Sampling:* 6 samples were collected over a lateral distance of about 1 mile. The analysis of entry (3) is based on 10 specimen measurements made on 3 of the 6 oriented samples collected; all 10 specimens have reversed polarity. Values for entry (4) are based on average sample-direction data obtained from Table 8, Ref. a; one divergent sample was discarded.
AUSTRALIA																
TASMANIAN DOLERITES																*Age:* The dolerite sills cut Late Triassic (possibly Early Jurassic) rocks and are involved in early Tertiary faulting; they therefore are Jurassic or Cretaceous. Assignment to the Jurassic is favored in Ref. b.
Surface samples	**5**	**42 S**	**147 E**	**325**	**−85**	**4**	**48**	**30**	**50 N**	**23 W**	**S**	**8**	**8**	a	*Irving* a Irving, 1956b b Irving, 1959	*Sampling:* 2 samples were collected at each of 30 sites covering an area of more than 9000 square miles. The mean site directions were used in the analysis. *Stability:* Stability is indicated by the application of Graham's conglomerate test to Tertiary breccia containing fragments of the dolerite.
Core samples	6	42 S	147 E	—	−85½	—	—	57	—	—	M	—	—	a	*Almond et al.* a Almond, Clegg. and Jaeger, 1956	*Sampling:* 3 cores were sampled over the depths 1–947 feet, 8–152 feet, and 500–636 feet, respectively. The cores were vertical, but azimuthally unoriented. *Other:* Since $I = 85½°$ is appropriate to latitude 81°, the pole should be roughtly 9° from the sampling area. *Reversals:* About one-fifth of the samples near the bottom of one core had reversed polarity.
	7	**42 S**	**147 E**	**—**	**−85½**	**—**	**—**	**57**	**42 N**	**33 W**	**M**	**>9**	**>9**	*		

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION H—*Continued*																
INDIA																
UPPER RAJMAHAL TRAPS	8	25 N	88 E	328	−64	—	—	33	13 N	70 W	S	—	—	a	*Clegg et al.* a Clegg, Radakrishnamurty, and Sahasrabudhe, 1958	*Age:* The lower Rajmahal traps are interbedded with sediments of Early to Middle Jurassic age (Krishnan, 1956, p. 273). The upper part of the traps are undated but petrologically similar to post-lower Cretaceous Deccan traps (Hobson, quoted in Pascoe, 1929, p. 146). Assignment to the Jurassic is favored in Ref. a *Sampling:* The samples came from the upper 250 feet of the traps. 33 oriented samples were collected from 3 quarries about 20 miles apart. Samples from 2 additional badly weathered outcrops had scattered directions of magnetization. Entry (9) is based on data scaled from Fig. 1, Ref. a. Entry (10) is based on the "secular-variation precision" $k' = 30$ and an estimate of the number of flows.
	9	25 N	88 E	327	−64	4	36	33	13 N	69 W	S	6½	5	*		
	10	**25 N**	**88 E**	**327**	**−64**	**10½**	***k'***	**6**	**13 N**	**69 W**	**S**	**17**	**13**	*		
AFRICA																
MADAGASCAR LAVAS AND DYKES	11	—	—	—	—	13½	32	5	58½N	162½W	S	13½	13½	*	*Roche et al.* a Roche, Cattala, and Boulanger, 1958 b Roche and Cattala, 1959	*Age:* Cited as Turonian stage of the Cretaceous *Sampling:* Entry (11) was obtained by a Fisher statistical analysis of 5 pole positions calculated from mean directions of magnetization of 5 sites (p. 2923, Ref. a). The 5 sites span a distance of 500 miles. Entry (13) was based on pole positions corresponding to the mean directions at 10 sites (p. 1050, Ref. b); these include (?) the data given in Ref. a. Entry (12) was found by grouping the results according to sampling area and averaging the results. *Stability:* Stability was checked by repeated measurements (Thellier's test) and by partial heat demagnetization.
	12	—	—	—	—	—	—	—	68 N	168 W	S	—	—	b		
	13	**—**	**—**	**—**	**—**	**9**	**31**	**10**	**66½N**	**163½W**	**S**	**9**	**9**	*		

SECTION H—*Continued*

JAPAN

Stability: Stability was checked by repeated measurements (Thellier's test) and by partial heat demagnetization.

SECTION H—*Continued*

JAPAN																
Inkstone Red Shales															*Nagata et al.* a Nagata, Akimoto, Shimizu, Kobayashi, and Kuno, 1959	*Age:* Cited as Middle to Early Cretaceous *Sampling:* 24, 15, and 23 oriented samples were collected at Okoti, Imoziya, and Okuhu, respectively. *Other:* The between-site dispersion is much greater than within sites. The value of α_{95} cited in entry (20) is therefore probably much too small. The value of k for entries (15), (17), and (19) was calculated by the approximate formula.
Okoti	14	—	—	81	54	4	—	24	—	—	S	—	—	a		
	15	34½N	131½E	81	54	4	51	24	25½N	164½W	S	5½	4	*		
Imoziya	16	—	—	50	46	4	—	15	—	—	S	—	—	a		
	17	34½N	131½E	50	46	4	82	15	47 N	143½W	S	5	3	*		
Ohuku	18	—	—	43	46	7	—	23	—	—	—	—	—	a		
	19	34½N	131½E	43	46	7	17	23	52½N	139½W	S	9	5½	*		
Combined	**20**	**—**	**—**	**58**	**50**	**2**	**—**	**62**	**42 N**	**153 W**	**S**	**3**	**2**	**a**		

SECTION I EOCENE

EUROPE																
Lundy Dykes	1	—	—	197	−59	—	—	8	—	—	—	—	—	a	*Blundell* a Blundell, 1957	*Age:* Not known but assigned to the early Tertiary (Ref. a) because the magnetic direction is similar to that for other lower Tertiary rocks from the British Isles *Sampling:* At least 2 samples were taken from each of 13 dikes and 2 dike contact rocks. 5 dikes and the 3 contact rocks were stable. Entry (2) was based on data from Table 1, Ref. a. *Stability:* Directions of magnetization were remeasured after 1 month and had not changed.
	2	**51 N**	**4½W**	**194**	**−59**	**7**	**64**	**8**	**75 N**	**129½E**	**N**	**10½**	**7½**	*		
Arran Dykes	**3**	**—**	**—**	**—**	**—**	**—**	**—**	**10**	**79 N**	**154 E**	**M**	**15**	**12**	**a**	*Leng* a Irving, 1959	*Age:* Cited as early Tertiary in Ref. a
Mull Intrusive Rocks	**4**	**56½N**	**6 W**	**½**	**58**	**14**	**20**	**7**	**72 N**	**173 E**	**M**	**21**	**15½**	*	*Vincens* a Vincenz, 1954 b Irving, 1959	*Age:* Early Tertiary, post Mull lavas and pre-Northwest Dikes *Sampling:* An average of about 9 samples was taken from each of 12 intrusive bodies. Only 7 of these bodies could be shown to possess a stable magnetization. The maximum separation of the sites was about 6 miles. Data for entry (4) were taken from Table I, Ref. a. *Reversals:* 2 of the 7 stable bodies showed reversed polarity.
	5	—	—	—	—	—	—	—	72 N	139 E	M	14	11	b		

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION I—*Continued*																
MULL LAVAS	6	56½N	6 W	137	−78	23½	—	16	—	—	—	—	—	b	*Bruckshaw and Vincens*	*Age:* Early Tertiary, probably Eocene (Ref. a) but may be Oligocene or even Miocene (Ref. b).
	7	**56½N**	**6 W**	**154½**	**−73½**	**17½**	**11**	**8**	**76½ N**	**74 W**	**N**	**31½**	**28½**	*	a Bruckshaw and Vincenz, 1954	*Sampling:* An average of about 7 samples was taken from each of 16 flows at 3 sites. 8 of the 16 flows could be shown to have stable magnetizations, and the data for entry (7) were for these stable flows, cited in Table I, Ref. a. Entry (6) is based on the data for both stable and unstable flows.
	8	—	—	—	—	—	—	8	82 N	71 W	N	31	28	c	b Hospers, 1955	
															c Irving, 1959	
ANTRIM BASALTS	9	55 N	6½W	194	−60	5½	—	24	—	—	N	—	—	a	*Hospers and Charlesworth*	*Age:* The interbasaltic zone covering these basalts is "probably Eocene or Oligocene but may be later" (Charlesworth, as reported in Ref. a). Simpson (reported in Ref. a) considers this zone to be late Miocene or early Pliocene.
	10	**55 N**	**6½W**	**194**	**−60**	**5½**	**31**	**24**	**73 N**	**135 E**	**N**	**8½**	**6½**	*	a Hospers, 1955	
	11	—	—	194	−60	—	—	—	75 N	118 E	N	—	—	b	b Creer, Irving, and Runcorn, 1954	*Sampling:* 6 flows at each of 3 sites and 5 at a fourth site were sampled; 2 to 4 samples were taken from each flow. The sites were about 30 miles apart (Ref. d).
	12	55 N	6 W	194	−60	5	—	—	74 N	133 E	—	8	6	c	c Irving, 1956a	
															d Hospers and Charlesworth, 1954	
ANTRIM BASALTS															*Wilson*	*Age:* Generally regarded as Eocene, but comparison of Scottish and Irish pollens suggests (Ref. a) an age as young as Miocene
Lower olivine basalts	13	—	—	173	−64½	—	—	—	—	—	—	—	—	a	a Wilson, 1959	*Sampling:* Ninety samples were collected from 57 igneous bodies at 47 sites spanning about 75 miles. The remanent magnetizations of one core from each sample and of the entire sample were both measured, and the mean of these two directions used in the statistical analysis. Three groups of basalt were studied: older "lower" olivine basalts (19 flows, entries 13 and 14); younger "middle" tholeiitic basalts (6 flows, entries 15 and 16); and intrusive rocks of unknown relative age (16 igneous bodies, entries 17 and 18). Entries (19) and (20) combine all these data. Statistical data were calculated from other statistical parameters reported in Ref. a.
	14	55 N	6 W	173	−64½	9½	13	19	80 N	162 W	N	15	12	*		
Middle tholeiitic basalts	15	—	—	206	−62	—	—	—	—	—	—	—	—	a		
	16	55 N	6 W	206	−62	16	18	6	69½ N	108 E	N	25	19	*		
Intrusive bodies	17	—	—	184	−63½	—	—	—	—	—	—	—	—	a		
	18	55 N	6 W	184	−63½	8½	20	16	79½ N	157 E	N	13	10½	*		
Combined	19	—	—	183	−64	—	—	—	—	—	—	—	—	a		
	20	**55 N**	**6 W**	**183**	**−64**	**6**	**16**	**41**	**80½ N**	**162 E**	**N**	**9½**	**7½**	*		

and 18). Entries (19) and (20) combine all these data. Statistical data were calculated from other statistical parameters reported in Ref. a.

NORTH AMERICA																
Siletz River Volcanic Rocks	21	—	—	70	55	7	—	8	37 N	50 W	M	—	—	a	*Cox*	*Age:* Early middle to early Eocene age established by interbedded fossiliferous sediments
	22	45 N	123½ W	70	55	7	50	8	37 N	49 W	M	10	7	*	a Cox, 1957	*Sampling:* 57 samples were taken from 8 flows over a lateral distance of 38 miles; the value of k was estimated by the approximate formula. *Stability:* Heat demagnetization and application of Graham's fold test decrease scatter. *Reversals:* 5 flows indicate normal (S) poles, and 3 indicate reversed (N) poles.
Laney Shale Member of Green River Formation	23	—	—	—	—	—	—	—	86 N	164 W	S	10	8	b	*Torreson et al.*	*Age:* Cited as Eocene in Ref. a
	24	41½N	109½W	353½	62½	6	30	19	85 N	170 W	S	9½	7½	*	a Torreson, Murphy, and Graham, 1949 b Irving, 1959	*Sampling:* 21 oriented samples were collected. 2 of these were too weakly magnetized to measure. Data for entry (24) were taken from Table 4, Ref. a. *Stability:* The circle of confidence includes the presen pole, thus there is no indication of stability.
Green River Formation (sediments)	25	39½N	108 W	345½	65	4½	168	7	77½ N	158 W	S	7	6	*	*Torreson et al.* a Torreson, Murphy, and Graham, 1949	*Age:* Cited as Eocene in Ref. a *Sampling:* 7 out of 9 oriented samples collected were strong enough to be measured. Data for entry (25) were taken from Table 4, Ref. a.
Wasatch Formation (sediments)	26	44½N	109 W	351½	63½	17	30	4	84½ N	180 E	S	26½	21	*	*Torreson et al.* a Torreson, Murphy, and Graham, 1949	*Age:* Cited as Eocene in Ref. a *Sampling:* 4 oriented samples were collected. Entry (26) is based on data from Table 4, Ref. a. *Stability:* No stability is indicated since circle of confidence includes present pole. *Other:* 5 additional samples from the Wasatch at Gardner, Colorado, showed random directions.
AUSTRALIA																
Older Volcanic Rocks of Victoria	27	38 S	145½E	17	−73	7	35	15	67 N	57 W	M	12	11	a	*Irving and Green* a Irving and Green, 1957	*Age:* Cited as early Tertiary and probably Eocene in Ref. a *Sampling:* 3 oriented samples at each of 15 sites were collected. The sites cover an area of approximately 5000 square miles. *Reversals:* 9 sites are normal (S poles), 4 sites are reversed (N poles), and 2 sites are mixed.

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION I—*Continued*																
TASMANIAN BASALTS	**28**	**42 S**	**147 E**	—	**−83**	—	—	**8**	**42 N**	**33 W**	**S**	**>13½**	**>13½**	*	*Almond et al.* a Almond, Clegg, and Jaeger, 1956	*Age:* Cited as early Miocene or Oligocene on basis of sediments associated with basalts (Ref. a); however, Banks (1958, personal communication) believes age may be Eocene. *Sampling:* 8 cores not oriented with respect to azimuth from 2 borings span a stratigraphic interval of 339 and 53 feet, respectively and penetrate a number of flows. The cores have an average inclination of 83°. *Other:* Since I = 83° corresponds to latitude 76½°, the pole should be roughly 13½° from the sampling locality. *Reversals:* One sample is reversely magnetized, but this core may have been inverted (Ref. a).
INDIA																
DECCAN TRAPS																
Undifferentiated	29	18 N	74 E	149	56	10	—	7	28 N	78 W	M	15	10	a	*Irving* a Irving, 1956a	*Age:* Generally considered Cretaceous to Eocene *Sampling:* 7 samples were collected at different levels at sites spanning some 200 miles. The data for the calculations of entry (30) were taken from the appendix of Ref. a. *Reversals:* 5 of the 7 samples showed reversed polarity (N poles) from the other 2.
	30	**18 N**	**74 E**	**149**	**56**	**13½**	**21**	**7**	**28 N**	**78 W**	**M**	**19½**	**14**	*		
Linga Area	31	22 N	79 E	164	48	—	—	195	—	—	—	—	—	a	*Clegg et al.* a Clegg, Deutsch, and Griffiths, 1956	*Sampling:* The 195 oriented samples or specimen measurements (?) were taken from 4 flows which were sampled over an area of some 50 square miles. The lowest of these flows is considered to be the bottom of the Deccan trap sequence. Values for α_{95} and k for entry (32) were calculated by the approximate formulas from α_{50} given in Ref. a.
	32	22 N	79 E	164	48	2	25	195	36½ N	83 W	N	2½	2	*		
Khandala Area	33	18½N	73½E	147	58	—	—	233	—	—	—	—	—	a		*Sampling:* 233 specimen measurements were made on 139 oriented samples from "numerous flows" collected at 20 sites. About 40 anomalous specimen measurements were not included. The flows are considered to be younger than those at Linga. Values of α_{95} and k for entry (34) were obtained in the manner outlined for entry (32) using data in Ref. a.
	34	18½N	73½E	147	58	3	9	233	25 N	78 W	N	4½	3½	*		
Linga and Khandala combined	35	—	—	155	53	—	—	—	28 N	85 W	N	—	—	a		*Sampling:* Average of Linga and Khandala results.

Linga and Khandala combined	35	—	—	155	53	—	—	—	28 N	85 W	N	—	—	a		*Sampling:* Average of Linga and Khandala results.
Kambatki Area	36	17½N	74 E	176	60	—	—	5	—	—	—	—	—	a	*Clegg et al. and Deutsch et al.*	*Sampling:* Of the 2 flows exposed at this area only 1 was stable, and the 5 oriented samples were from this flow. This flow is considered (Ref. b) to be near the bottom of the sequence. Values of k and α_{95} for entry (37) were obtained in the manner outlined for entry (32) using data in Ref. a. Since only one point in time was sampled, α_{95} as a measure of the average field direction should be larger.
	37	17½N	74 E	176	60	5	150	5	31 N	102 W	N	8	6	*	a Clegg, Deutsch, Everitt, and Stubbs, 1957	
Igatpuri Area	38	19½N	73½E	161	51	—	—	—	—	—	—	—	—	b	b Deutsch, Radakrishnamurty and Sahasrabudhe, 1959	*Sampling:* 4 samples were taken from each of 4 flows (Ref. b). One flow was unstable.
	39	19½N	73½E	161	51	—	—	—	35 N	86½W	N	—	—	*		
Nipani Area, Lower flows	40	16½N	74½E	170	57	—	—	44	—	—	—	—	—	a		*Sampling:* One flow was sampled at two sites (Ref. b). 4 anomalous specimens from 1 sample were excluded. Values of α_{95} and k for entry (42) were obtained as outlined for entry (32) using data in Table 2 of Ref. a.
	41	16½N	74½E	168	60	—	—	44	—	—	—	—	—	b		
	42	16½N	74½E	168	60	4	29	44	31½N	95 W	N	6	4½	*		
Nipani Area, Upper flows	43	16½N	74½E	340	−30	—	—	59	—	—	—	—	—	a		*Sampling:* At least 3 distinct flows spanning a stratigraphic thickness of 200 feet were sampled at 5 localities. These flows probably overlie the lower Nipani flows. Values of α_{95} and k for entry (45) were obtained as outlined for entry (32) using data in Ref. b.
	44	16½N	74½E	338	−32	—	—	74	—	—	—	—	—	b		
	45	16½N	74½E	338	−32	4	14	74	50 N	71½W	S	4½	2½	*		
Amba Area, Lower flows	46	17 N	74 E	141	59	—	—	54	—	—	—	—	—	a		*Sampling:* At least 5 flows were sampled at 8 sites in the Amba area; the results from the 5 lowest sites are combined here. Several samples with widely divergent directions were not used in the statistical analysis. Values of α_{95} and k for entry (48) were obtained as outlined for entry (32) using data in Ref. b.
	47	17 N	74 E	144	60	—	—	109	—	—	—	—	—	b		
	48	17 N	74 E	144	60	4	10	109	23 N	77 W	N	6	4½	*		
Amba Area, Upper flows	49	17 N	74 E	335	−23	—	—	34	—	—	—	—	—	a		*Sampling:* The results from the highest 3 of the 8 Amba sites are combined here; at least 5 flows were sampled at the 8 Amba sites. Values of α_{95} and k for entry (51) were obtained as outlined for entry (32) using data in Ref. b.
	50	17 N	74 E	335	−26	—	—	54	—	—	—	—	—	b		
	51	17 N	74 E	335	−26	7	8	54	50½N	66 W	S	7½	4	*		

TABLE 1.—*Continued*

Rocks Sampled	No.	Locality		Magnetic Direction					Pole Position						References	Remarks
		Lat.	Long.	Decl.	Incl.	α_{95}	k	N	Lat.	Long.	P	δm	δp	S		
SECTION I—*Continued*																
Pavagadh Area, Lower flows	52	22½N	71½E	348	−12	—	—	41	—	—	—	—	—	a		*Sampling:* 8 basic flows lying below an elevation of 2425 feet on Mt. Pavagadh were sampled. Values of α_{95} and k for entry (54) were obtained as outlined for entry (32) using data in Ref. b.
	53	22½N	71½E	351	−16	—	—	69	—	—	—	—	—	b		
	54	22½N	71½E	351	−16	7	5	69	58 N	91 W	S	7½	4	•		
Pavagadh Area, acid tuffs	55	22½N	71½E	357	15	—	—	17	—	—	—	—	—	a		*Sampling:* 15 specimen measurements were made on 8 samples of acid tuff collected at an elevation above 2490 feet on Mt. Pavagadh. They are post Deccan traps and may be post Eocene. Values of α_{95} and k for entry (57) were calculated as outlined for entry (32) using data in Ref. b.
	56	22½N	71½E	355	17	—	—	15	—	—	—	-	-	b		*Other:* Associated rhyolite flows show random directions, and laboratory tests indicate that these flows are unstable.
	57	**22½N**	**71½E**	**355**	**17**	**7**	**26**	**15**	**75½N**	**88 W**	**S**	**7½**	**4**	•		
UPPER DECCAN TRAPS AVERAGE	**58**	—	—	—	—	**13½**	**82**	**3**	**53 N**	**75½W**	**S**	**13½**	**13½**	•		*Sampling:* These values were computed by applying Fisher statistics to the upper Amba (51), upper Nipani (45), and lower Pavagadh (54) pole positions. There is no independent stratigraphic evidence correlating all these units, but the occurrence at several localities of normally magnetized flows overlying reversely magnetized flows has been interpreted (Ref. b) as supporting the correlation of all the normal flows and all the reversed flows.
LOWER DECCAN TRAPS AVERAGE	**59**	—	—	—	—	**8½**	**65**	**6**	**30½N**	**87 W**	**N**	**8½**	**8½**	•		*Sampling:* These values were computed by applying Fisher statistics to the lower Amba (48), lower Nipani (42), Igatpuri (39), Kambatki (37), Khandala (34), and Linga (32) pole positions. See also previous note.

Discussion of Results

General Statement

An attempt is here made to separate a discussion of the paleomagnetic data from a discussion of the conclusions that have been drawn from these studies. Individual paleomagnetic studies are briefly described first, emphasizing tests for stability and paleomagnetic applicability. All the virtual geomagnetic poles printed in boldface type in Table 1 are shown in the figures that accompany this section, and, since the reliability of the determinations varies widely, frequent reference to the data tables is recommended. The symbols used in the illustrations are explained in Table 2.

The sequence we have used for presenting the data places the post-Eocene first; this was done because these results are relatively simple and furnish a norm for comparison with the older results. The rest of the stratigraphic column is then discussed in sequence, beginning with the Precambrian. The section concludes with a discussion of the applications of paleomagnetic results to hypotheses such as polar wandering and continental drift.

Table 2.—Key to Figures 17, 20, 22–31, 35

ТАБЛИЦА 2.—УСЛОВНЫЕ ОБОЗНАЧЕНИЯ К РИСУНКАМ 17, 20, 22–32 и 35

VIRTUAL GEOMAGNETIC POLES

ИСТИННЫЕ ГЕОМАГНИТНЫЕ ПОЛЮСЫ

- ○ EUROPEAN
  ЕВРОПЕЙСКИЕ
- □ NORTH AMERICAN
  СЕВЕРО-АМЕРИКАНСКИЕ
- △ AUSTRALIAN
  АВСТРАЛИЙСКИЕ
- ▽ INDIAN
  ИНДИЙСКИЕ
- ▭ SOUTH AMERICAN
  ЮЖНО-АМЕРИКАНСКИЕ
- ◇ AFRICAN
  АФРИКАНСКИЕ
- ▯ ASIAN
  АЗИАТСКИЕ

POLARITY SYMBOLISM (APPLIES TO ALL SHAPES ABOVE AND NOT MERELY TO CIRCLES)

СИМВОЛЫ ПОЛЯРНОСТИ /ОТНОСЯТСЯ НЕ ТОЛЬКО К КРУЖКАМ, НО И КО ВСЕМ ВЫШЕПРИВЕДЕННЫМ УСЛОВНЫМ ОБОЗНАЧЕНИЯМ/

- ● SOUTH MAGNETIC POLES ("NORMAL" POLARITY).
  ЮЖНЫЕ МАГНИТНЫЕ ПОЛЮСЫ/"НОРМАЛЬНАЯ" ПОЛЯРНОСТЬ/
- ○ NORTH MAGNETIC POLES ("REVERSED" POLARITY).
  СЕВЕРНЫЕ МАГНИТНЫЕ ПОЛЮСЫ/"ОБРАТНАЯ" ПОЛЯРНОСТЬ/
- ◑ POLES FROM MEASUREMENTS WITH MIXED POLARITY.
  ПОЛЮСЫ, ПОЛОЖЕНИЕ КОТОРЫХ ВЫСЧИТАНО ПО ИЗМЕРЕНИЯМ СМЕШАННОЙ ПОЛЯРНОСТИ
- 95 PER CENT CONFIDENCE LIMITS
  ГРАНИЦА 95% ТОЧНОСТИ

Post Eocene

Late Pleistocene and Recent.—The virtual geomagnetic pole positions for the data from this period are given in Section A of Table 1 and are shown graphically in Figure 17. The total sampling area represented is large, since two of the determinations are from North America, two are from Asia, and the remaining nine

European studies were made in areas as far apart as Iceland, Sicily, and Northern Sweden. The Sicilian measurements (**A 2**) (Chevallier, 1925, p. 146–158) were made on 11 lava flows 1600, but the earlier Roman measurements have westerly declinations.

Measurements on 150 samples from the Ångerman River varves (Griffiths, 1955, p. 108)

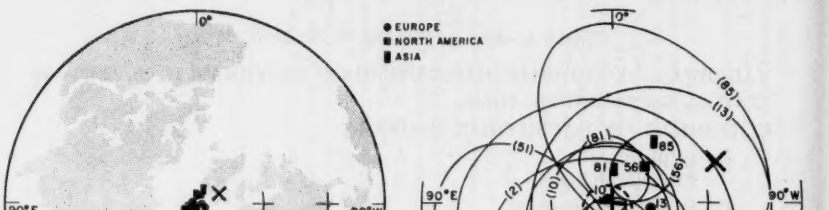


FIGURE 17.—LATE PLEISTOCENE AND RECENT VIRTUAL GEOMAGNETIC POLES (TABLE 1, SEC. A)

Mean poles and 95 per cent confidence intervals in region north of 70° N. Lat. are shown on the right. Heavy dashed circle is "α_{95}" calculated from mean pole positions but has no rigorous statistical significance. Numbers refer to entries in section A, Table 1.

from Mt. Etna extruded over a period of about 2300 years, the last one in 1911. The last two flows have average directions agreeing with nearby observatory measurements of the field, but there is no evidence of stability, other than the fact that the older flows have directions different from the present field direction. The Prästmon varves (**A 6**) (Bancroft, 1951) cover the period from 0 to A.D. 1000 and come from an area some 25 degrees north of the Sicilian lavas.

Measurements of the remanent magnetism of kilns from Carthage have been reported by Thellier and Thellier (1951, p. 1477) (**A 4**). Two of the kilns were last fired in 146 B.C., and the third in 300 A.D., so that only two points in time are represented, and no statistical calculations can be made. Measurements on 14 archaeological specimens of fired clay (**A 7**) have been reported by Cook and Belshé (1958, p. 174). These specimens were all collected from Great Britain and are from the Roman period (first–fourth centuries) and the Middle Ages (twelfth, thirteenth, and fifteenth centuries). The measurements from the Middle Ages show easterly declinations, in agreement with direct observatory measurements made in London in collected at two localities a few kilometers apart were averaged into 29 groups each representing about 100 years; the mean value for each group was then used in the statistical analysis to compute pole **A 8**. The period covered is from 1100 B.C. to A.D. 750.

Brynjólfsson (1957, p. 252) has measured the direction of magnetization in 21 post-glacial lava flows from Iceland (**A 10**). Specimens from these flows, which cover the period from 3400 B.C. to A.D. 1950, were partially demagnetized in A.C. fields of 140 oersted before measurement, indicating that the reported directions were determined from magnetic components having high coercivities. Post-glacial lava flows from Iceland have also been measured by Hospers (1955, p. 63) (**A 12**). The consistency of both Bryjólfsson's and Hosper's measurements is good, k = 36 and k = 34, respectively.

Granar (1959, p. 27) sampled 10 sections of Swedish glacial and post-glacial varves over a lateral extent of some 800 km (**A 13**). Measurements on 10 upper Pleistocene lava flows from France (Roche, 1958, p. 3365) give the average pole numbered **A 15**. (Data necessary for a Fisher analysis were not available.)

Some of the first remanent magnetic measure-

ments made on rocks from North America were those of Johnson and others (1948, p. 366) on glacial varves from several localities in New England (**A 51**). These measurements were grouped into sets representing about 1000 years each, covering the period from 13,000 to 7000 B.C. The average virtual geomagnetic pole is displaced from the geographic pole, and inclination errors such as those observed in artificially deposited varves (King, 1955, p. 121) would cause such a pole displacement in the observed direction, as Johnson and others (1948, p. 358) suggested. The magnetization direction in the varves thus probably does not represent the field direction at the time they were deposited, even though a fold test demonstrated the magnetic stability of at least some New England varves (Graham, 1949, p. 137–143).

The other measurements from North America are those on the Neroly formation of California (**A 56**) (Doell, 1956, p. 158). Although the rocks were deposited much earlier, it was shown that the magnetization was acquired after folding of post early Pleistocene age because application of Graham's fold test causes considerable scatter in the directions (the *in situ* directions are consistent). Partial demagnetization by heating to 100°C and cooling in zero field did not alter the magnetization, suggesting stability.

Measurements on 11 lava flows of Pleistocene to Recent age from Japan are given by Kumagai and others (1950, p. 62) (**A 85**). Watanabe (1958, p. 383–384) has reported magnetic directions for 45 archaeological fired clays and also two historically dated lava flows from Japan (**A 81**). These measurements cover the periods from 300 to 1800 A.D. and from about 5600 to 4400 B.C.; thus, some 2700 years has been sampled.

Several other groups of measurements which contain rocks from both this and preceding Tertiary time are discussed in the section on Oligocene through early Pleistocene measurements.

In summary, each of the 13 average geomagnetic poles calculated for late Pleistocene and Recent time were calculated from sets of samples that were probably magnetized over a period of several thousand years. (One possible exception is the Neroly formation (**A 56**); the magnetic history of this rock is somewhat obscure.) These poles are very tightly grouped, and a comparison with Figure 13 indicates, in fact, a much higher precision than for the virtual geomagnetic poles calculated from present field directions. This initially surprising result is easily explained, however, because each of the 13 paleomagnetically determined poles represents an average of several points in time. The nondipole components, which cause scatter in the present field, apparently have tended to cancel each other out. These paleomagnetic results are of great significance, since they clearly indicate that the basic configuration of the earth's field was that of a dipole with a fixed axis during the time these groups of rocks were magnetized. The tight grouping of virtual geomagnetic poles would not occur if the earth's field were not dipolar, nor if the average dipolar axis had moved between the times the different groups were magnetized.

In discussing historic observations of the earth's field we noted that there is little direct evidence that the earth's inclined dipolar axis has moved. If this axis had also been in its present inclined position throughout late Pleistocene and Recent time, we should expect to find the poles in Figure 17 grouped about the present geomagnetic pole rather than about the earth's *geographic* pole. Values of α_{95} should also be considered in evaluating the significance of this grouping, and from the correlation diagram (Fig. 18) it can be seen that two ovals of confidence encircle both poles, two ovals (from the Neroly formation and the New England varves) encircle neither pole, no oval includes the geomagnetic but not the geographic pole, and seven include the geographic pole but exclude the geomagnetic pole. There is thus very strong paleomagnetic evidence that throughout late Pleistocene and Recent time the earth's field, when averaged over a few thousand years, has been that of a dipole parallel to the present axis of rotation.

The reversal problem.—In contrast with the late Pleistocene and Recent results, about half the older post-Eocene rocks have magnetizations about 180° removed from the present direction of the earth's field. Before considering these results in detail, it is necessary, therefore, to consider briefly the problem of reversals. As noted before there are two interpretations of this phenomenon, one that the earth's field periodically reverses its polarity or, alternatively, that the rocks having reversed polarity possess a self-reversal mechanism. Although the problem is not completely resolved, it appears to us that some reversals are due to self-reversal and others to a reversal of the field. The investigation of the Haruna dacite by Uyeda (1958, p. 29–48) clearly shows that rocks can be self-reversing, and the correlation between inclina-

tion and mineralogy found by Balsley and Buddington (1954, p. 180) in metamorphic rocks strongly suggests a self-reversal. However, in thick sequences of alternating normal and reversed lava flows such as those in Iceland the reversals are not associated with changes in lithology (Einarsson, 1957, p. 233), and the PTRM curves for the reversed flows give no indication of a self-reversal mechanism (Hospers, 1953–1954, p. 487). Evidence given by Roche (1953, p. 108) also favors a reversal of the field; he found that four subjacent clays baked by reversely magnetized lava flows were also reversely magnetized. It seems improbable that in all four cases the subjacent clays as well as the lavas themselves would possess a self-reversal property. Einarsson and Sigurgeirsson (1955, p. 892) have also found numerous similar baked zones in Iceland; the baked rocks all have the same direction of magnetization as the baking flow (Fig. 19).

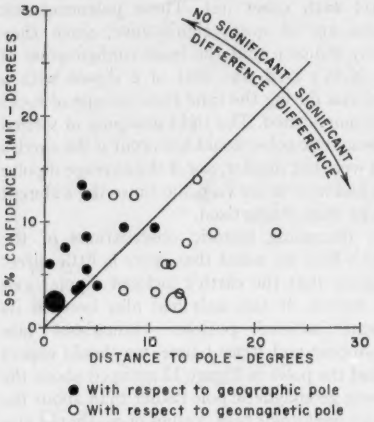


FIGURE 18.—CORRELATION DIAGRAM FOR LATE PLEISTOCENE AND RECENT POLES

Two points are plotted for each virtual geomagnetic pole (VGP); the abscissae of the two points are the lengths of the lines from the VGP to the geographic and geomagnetic poles respectively, and the ordinates are the semiaxes of the oval of confidence along these two lines. A point in the lower "significant difference" field thus indicates the pole lies outside the confidence interval. Large circles are for mean VGP position shown by heavy dashes in Figure 17.

The youngest reversely magnetized rocks are early Pleistocene or very late Pliocene in all the regions where reversely magnetized rocks have been found, including Japan (Nagata and others, 1957, p. 32), Iceland (Einarsson and Sigurgeirsson, 1955, p. 892; Hospers, 1953–1954, p. 475), France (Roche, 1953, p. 109; 1956, p. 814), and Russia (Khramov, 1957, p. 851). Although the exact ages of the rocks in most of these studies are subject to some uncertainty,

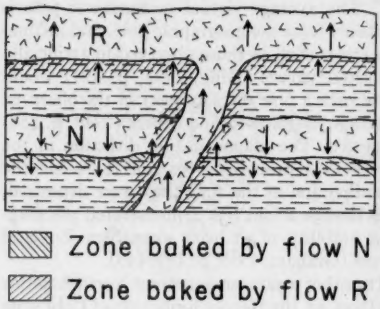


FIGURE 19.—MAGNETIC DIRECTIONS IN BAKED ZONES

the stratigraphic control is adequate to indicate that none of the late Pleistocene or Recent rocks which have been studied paleomagnetically are reversely magnetized. (The Haruna dacite samples were collected from a tuff for which there are no paleomagnetic data, but a young flow of this rock would presumably be reversed.) The last appearance of reversely magnetized rocks at about the same time in such widely separated areas suggests field reversal.

Oligocene through early Pleistocene.—The virtual geomagnetic poles from paleomagnetic studies on Oligocene through early Pleistocene rocks are also given in Section A of Table 1. Thirteen of the 28 studies are from Europe, and the remainder from other continents. The most notable feature of these data in comparison with the late Pleistocene and Recent data is the large number of reversed magnetizations. Nine are of mixed polarity, 11 are consistently reversed, and 7 are entirely of normal polarity (Fig. 20).

Results from Icelandic lava flows of early Quaternary and Miocene age are given by Hospers (1955, p. 65, 68) and of Pliocene and Pleistocene age by Sigurgeirsson (1957, p. 243). The Quaternary flows studied by Hospers (**A 17**) all show reversed polarity, and although the precision is not high ($k = 9$) the reversed nature of the magnetization indicates stability. The Miocene flows (102 in number) (**A 34**) show about equal numbers of normal and reversed polarizations, occurring in four alternately reversed and normal zones. The consistency-of-

reversals test indicates stability, and, because of the large number of flows measured, α_{95} is small despite the low value of k. Sigurgeirsson's studies, in which the samples were "cleaned" in A.C. fields of 110 oersted, resulted in much larger values for k. The mean pole for the reversely magnetized flows (**A 25**) is not significantly different from the geographic pole, but that for the normally magnetized flows (**A 23**) is, an effect which Sigurgeirsson (1957, p. 243) attributes to the small number of flows sampled in one part of the normal sequence.

Five determinations of the remanent magnetization of lava flows and intrusives from France, ranging in age from Oligocene to early Quaternary, have been reported by Roche (1958, p. 3365; 1951, p. 1133; 1950b, p. 1604; 1950a, p. 114). Six flows of Pliocene and Pleistocene age (**A 21**) have an average direction not significantly different from the geographic pole. However, five flows of Miocene and Pliocene age (**A 31**) and nine determinations on Oligocene intrusives (**A 48**) have circles of confidence that do not include the present geographic pole. Statistical data for the Limagne basalt (**A 46**) and eight flows of early Quaternary age (**A 19**) were not available. All the samples studied in these five determinations had reversed polarity.

Khramov (1957, p. 850) (**A 27** and **A 29**) studied 650 oriented samples of Pliocene and Pleistocene sediments from the Chelekan Peninsula. The samples show both normal and reversed polarity and indicate pole positions relatively far from the geographic pole. Because statistical data were not given a confidence interval could not be calculated.

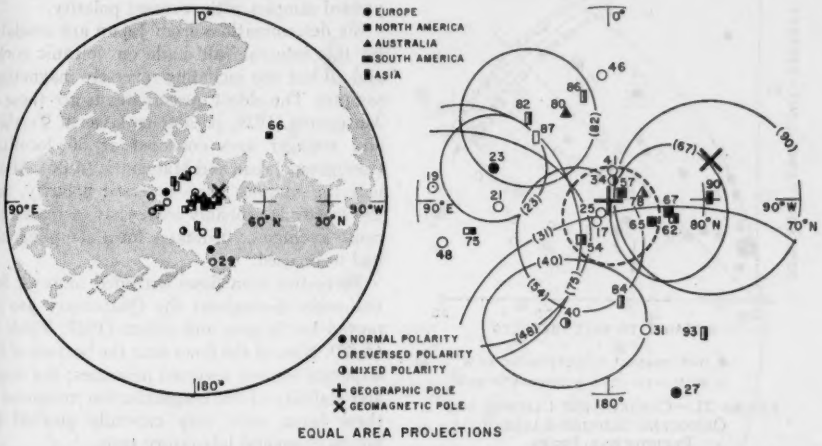


FIGURE 20.—OLIGOCENE THROUGH EARLY PLEISTOCENE VIRTUAL GEOMAGNETIC POLES (TABLE 1, SEC. A)

Mean poles in region north of 70° N. Lat. are shown on the right. Ovals of confidence are shown for only those poles significantly different from the geographic pole. Heavy dashed circle is "α_{95}" calculated from mean pole positions but has no rigorous significance. Numbers refer to those in section A, Table 1.

The circle of confidence calculated from the directions of magnetization of 42 lava flows from the Vogelsberg (**A 40**) (Angenheister, 1956, p. 190–191) includes neither the present geographic pole nor the geomagnetic pole. Thirteen of the flows have reversed polarity, and, treated separately, the reversed and normal flows do not have exactly the same average direction, indicating the presence of a secondary component of magnetization by the consistency-of-reversals test.

Bruckshaw and Robertson (1949, p. 316) found that all the samples from the North West Dikes of England (**A 41**) had directions of magnetization roughly opposed to the present field direction.

In contrast with the European measurements which, except for the Russian results, were all made on igneous rocks, most of the determinations from North America were made on sedimentary rocks. Twenty three oriented samples from the Ellensburg formation of late Miocene and early Pliocene age and 21 from the Arikaree formation of Miocene age studied by Torreson and others (1949, p. 125) yield poles **A 65** and **A 66** respectively. The poles are not significantly

different from the geomagnetic pole, and there is no indication of stability. Torreson and others (1949, p. 125) also measured 13 oriented samples from the Payette formation of Miocene and

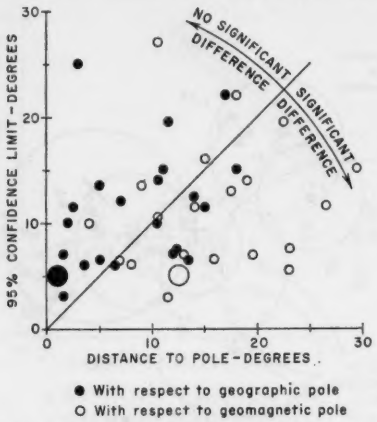


FIGURE 21.—CORRELATION DIAGRAM FOR OLIGOCENE THROUGH EARLY PLEISTOCENE POLES
See caption for Figure 18

Pliocene(?) age (**A 57**). These are all normally magnetized and have an unusually high precision parameter (k = 258). Collinson and Runcorn (1960) measured 24 samples of the Duchesne River formation from Utah (**A 67**). Like the other Tertiary sediments from North America these are all normally polarized. Du Bois (1959a, p. 1618) measured 46 samples collected from lava flows in northwestern Canada at four very widely separated localities (**A 54**). The direction of magnetization at two of the localities was reversed. The Columbia River basalt of Miocene age was extensively sampled by Campbell and Runcorn (1956, p. 450) (**A 62**). Twenty nine of the 73 flows sampled had reversed polarity.

Ten Quaternary lava flows from South America (**A 75**) have been studied by Creer (1958, p. 381). Some flows are normally, and some reversely polarized. The samples were all partially demagnetized in fields of 250 oersted A.C. before measurement so that the magnetizations reported are those of the higher coercive force magnetic constituents.

Two measurements are available from the southwest Pacific, one on the New Zealand ignimbrites (**A 80**) (Hatherton, 1954, p. 429), and the other on the Newer volcanics of Victoria (**A 78**) (Irving and Green, 1957, p. 351). Data necessary for computing Fisher statistics for the Pliocene ignimbrites were not available. The Newer volcanics of Victoria, sampled at 32 widely separated sites, range in age from Pliocene to Recent; 16 of these sites yielded samples with reversed polarity.

Six determinations from Japan are available for this interval—all made on volcanic rocks, and all but one including reversely magnetized samples. The oldest measurements are those of Matuyama (1929, p. 204) on lavas of Tertiary and younger ages collected at 36 localities throughout Japan and Manchuria (**A 84**). About half the samples have reversed polarity, and the scatter is probably somewhat increased because average co-ordinates for a sampling site had to be used.

Forty-two lava flows extruded more or less uniformly throughout the Quaternary are reported by Nagata and others (1957, Table 2) (**A 82**). Nine of the flows near the bottom of the sequence showed reversed polarities; the origin and stability of the magnetization measured in these lavas were very carefully studied by means of several laboratory tests.

Of the seven lava flows of late Tertiary age described by Kumagai and others (1950, p. 62) (**A 86**) only one has reversed polarity. One reversely magnetized lava flow of early Pleistocene age from Kawajiri was measured by Asami (1954a) giving pole **A 87**.

Two dolerite sheets, an andesite sheet, and four andesite lavas of Miocene age and two andesite lavas and two basalt lavas of Pliocene age have been studied by Nagata and others (1959, p. 380–381). The Pliocene flows (**A 90**) all have normal polarities, but three of the Miocene units (**A 93**) have reversed polarities.

In addition to the above studies, six lava flows stratigraphically between normally and reversely magnetized flows have intermediate directions of magnetization (Momose, 1958, p. 18); the data in this preliminary description are insufficient for the calculation of confidence limits.

Except for the occurrence of reversals and somewhat greater scatter in the virtual geomagnetic poles, the results from Oligocene through early Pleistocene time are very similar to those for the late Pleistocene and Recent. Only those ovals of confidence which do not encircle the present geographic pole are plotted in Figure 20; Figures 20 and 21 show that the poles farthest from the geographic pole generally have the largest ovals of confidence. Some of the

ovals of confidence which do not encircle the geographic pole appear, for statistical reasons, to be too small (Table 1). As may be seen in the correlation diagram, the poles definitely group about the geographic pole rather than the geomagnetic pole, and it appears that, within somewhat broader limits than indicated for the late Pleistocene and Recent, the earth's average magnetic field throughout post-Eocene time was that of a dipole parallel to the present axis of rotation.

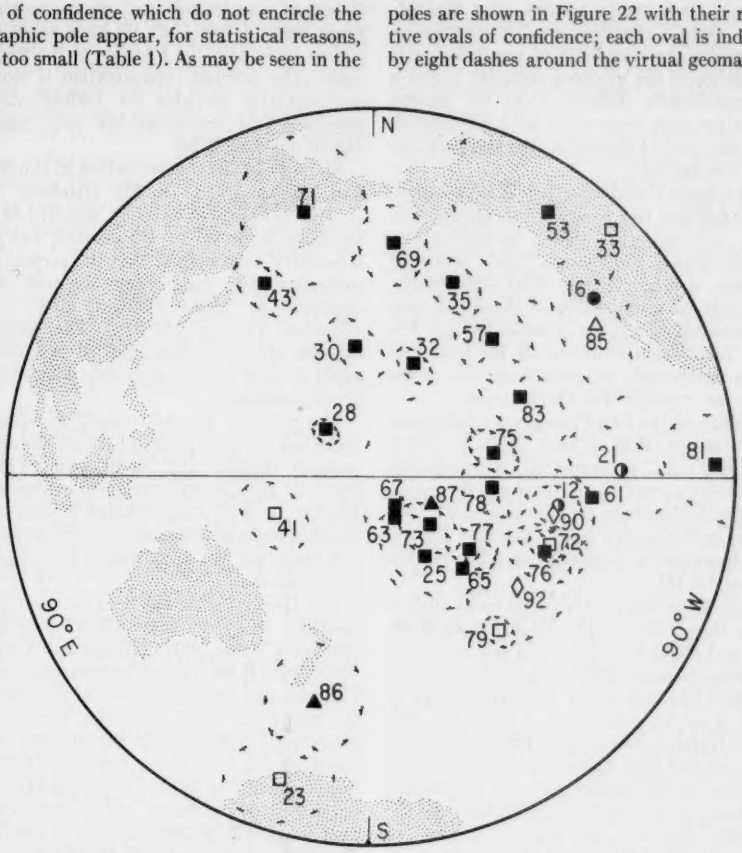


FIGURE 22.—PRECAMBRIAN VIRTUAL GEOMAGNETIC POLES (TABLE 1, SEC. B)

Precambrian

In striking contrast with results from the upper Tertiary, virtual geomagnetic poles calculated from all reported remanent magnetizations of rocks of Precambrian age are significantly different from the present geographic and geomagnetic poles. These virtual geomagnetic poles are shown in Figure 22 with their respective ovals of confidence; each oval is indicated by eight dashes around the virtual geomagnetic pole, a device introduced to avoid giving undue weight to data with large confidence intervals.

Two paleomagnetic studies have been made on British rocks of Precambrian age. From a paleomagnetic point of view the Torridonian sandstone (**B 12** and **B 16**) (Irving and Runcorn, 1957, p. 88) has been studied as carefully as any formation. Four hundred oriented samples were collected over a wide area and through a thick stratigraphic sequence, and the field tests indicating magnetic stability are very impressive. In conglomerates containing very fine-grained Torridonian pebbles, the directions of magnetization within pebbles are uniform, whereas those between pebbles are random; directions of magnetization in the beds are widely scattered before correcting for Caledonian

folding but are nearly parallel after the correction. Only a relatively small proportion of the samples show evidence of instability, and the balance of the evidence certainly favors a field significantly different from the present field at the time these rocks were magnetized, which was before Caledonian time and probably in the Precambrian.

In the upper Torridonian (pole **B 12**) magnetizations fall into two groups which are approximately reversed with respect to each other; since the axes of the "normal" and "reversed" directions of magnetization differ significantly, the statistical combination of these two may have resulted in too low a value for α_{95}. The lower Torridonian sequence (**B 16**) has a direction significantly different from that of the upper, and in contrast has no reversals.

Stability of the late Precambrian Longmyndian formation (**B 21**) (Creer, 1957b, p. 126) is suggested by A.C. demagnetization experiments and Thellier's test (that is, there was no change in magnetic directions on remeasurement after storage in the earth's field). The pole lies between those for the upper and lower Torridonian (**B 12** and **B 16**).

Of the Canadian Keweenawan rocks studied by Du Bois (1957, p. 178) (**B 23, B 25, B 28, B 30,** and **B 32**) only the Freda sandstone and the Nonesuch shale (**B 28**) are folded and satisfy Graham's fold test. These data were shown in Figure 6 as an example of a fold test. The Copper Harbour formation (**B 30**) includes both sediments and lava flows; the sediments have an inclination 10° smaller than the intercalated lava flows, which is much smaller than the 63° difference between the mean direction of remanent magnetization and the present direction of the earth's field. Random directions of magnetization in pebbles of Copper Harbour lava in conglomerates also suggest stability. Two sedimentary formations from Newfoundland studied by Nairn and others (1959, p. 596) give the pole positions **B 81** and **B 83**.

The gabbro intrusives from southeastern Canada (**B 35, B 41, B 43**) (Hood, 1958) show some evidence of metamorphism. The gabbro is locally gneissic, and laboratory results strongly suggest susceptibility anisotropy so that these measurements may not be suitable for paleomagnetic purposes, even though A.C. demagnetization usually decreased the scatter in directions. A stronger case exists for the stability of magnetization of the Sudbury intrusive. The mean directions of magnetization of sets of samples collected on the north and south rims of this body lie on a small circle having, as its axis, the fold axis of the Sudbury basin. The simplest interpretation is that the magnetization predates the folding; the appropriate fold correction has been made in calculating pole **B 53**.

Stability of the magnetization of the Michigan diabase dikes (**B 33**) (Graham, 1953, p. 246) is indicated by their stability in A.C. magnetic fields up to 493 oersted, but other laboratory tests suggest that the present magnetization may have been acquired during chemical or phase changes after cooling (Graham, 1953, p. 252–254). The confidence interval about the pole position is artificially small because at most only two points in time are represented.

The pole for the Adirondack metamorphic rocks has not been plotted because the original workers (Balsley and Buddington, 1958, p. 790–792) have given convincing evidence that the directions of magnetization in these metamorphic rocks are controlled by the mineralogy and hence are not applicable to paleomagnetic interpretations.

The Hakatai shale has been sampled at two localities about 1° apart (**B 57** and **B 63**) (Runcorn, 1956a, 309; Collinson and Rundorn, 1960). Pole **B 65** is based on measurements by Collinson and Runcorn (1960) of samples from the underlying Bass limestone, which has a gradational contact with the Hakatai. Combined results from these two units based on measurements by Doell (1955a) (**B 61**) have a large oval of confidence which intersects the oval for pole **B 57** but not that of pole **B 65**. The Shinumo quartzite (**B 67**) (Collinson and Runcorn, 1960) overlies the Hakatai shale. The pole positions for these related formations, while scattered, show some measure of consistency. However, the differences emphasize the need for extensive sampling before pole positions can legitimately be regarded as representative of a given continent and geologic period.

The results for flat-lying beds (**B 69**) from the Hazel formation (Howell and others, 1958, p. 291) and for dipping beds after correction for dip (**B 71**) are not significantly different in mean direction, but differ considerably in amounts of scatter ($k = 35$ for flat lying, $k = 3$ for tilted beds), suggesting the presence of an unstable component of magnetization. The rocks may be slightly metamorphosed and hence unsuited for paleomagnetic applications.

The Belt series (**B 72, B 73, B 75, B 76, B 77, B 78,** and **B 79**) has been sampled over a wide region by Collinson and Runcorn (1960). The virtual geomagnetic poles from this series form a remarkably tight group in the vicinity of 10° S. and 150° W., and it would be very difficult to regard this grouping near the present equator as random. The mean directions of magnetization at two sites (**B 73** and **B 85**) are approximately reversed with respect to the others, satisfying the consistency-of-reversals stability test.

Results are available from three Australian formations (Irving and Green, 1958, p. 66): the Buldiva quartzite (**B 85**), the Nullagine lavas (**B 86**), and the Edith River volcanics (**B 87**) which are probably younger than the Nullagine lavas.

The Pilansberg dikes (**B 90**) were sampled at widely spaced localities and show excellent consistency of magnetizations (Gough, 1956, p. 206). Moreover, stability is suggested by the small changes of magnetization in alternating fields of 100 oersted. The Bushveld gabbro (**B 92**) was also sampled over a wide area (Gough and van Niekerk, 1959, p. 131), and its stability of magnetization is suggested by a considerable reduction in scatter on correcting for geologic dip, as inferred from pseudostratification in the gabbro. If a preferred orientation or layering of ferromagnetic grains accompanies the stratification, however, the original direction of TRM may not have been parallel to the applied field; Graham's fold test does not eliminate this possibility, and measurements of susceptibility anisotropy would be desirable (Girdler, 1959). Recently determined radio-isotope ages for these two bodies are 1290 m.y. for the Pilansberg (Schreiner, 1958, p. 1330) and 2000 m.y. for the Bushveld (Gough and van Niekerk, 1959, p. 127). In view of the large difference in age, the small distance between the poles (18°) is remarkable.

In summary, all the paleomagnetically studied Precambrian formations are magnetized in directions significantly different from that of the present field. Although there are grounds for reasonable doubt about the paleomagnetic applicability of some of the formations described, many of them satisfy the classic tests for stability. These tests are Graham's fold and conglomerate tests, Thellier's test for change of direction in the laboratory, stability of magnetization in A.C. fields, concurrent results from igneous and sedimentary rocks, consistency of directions different from that of the present field over wide sampling areas, and consistency of reversals. The balance of the evidence strongly suggests that the earth's field was not parallel to its present direction throughout Precambrian time.

The distribution and polarity of the virtual geomagnetic poles calculated for Precambrian formations from all continents do not appear to be entirely random. Most of the poles fall in a region covering about one-third of the hemisphere, and a concentration appears near the equator in the vicinity of 160° W. Some of the poles considerably removed from this region, however, are based on excellent data (for example, poles **B 16** and **B 28**).

Early Paleozoic

Because of the scarcity of results from the Cambrian through Devonian formations, they are grouped together, and the virtual geomagnetic poles are shown in Figure 23. One Cambrian formation from Europe has been studied, the Caerbwdy sandstone from South Wales (**C 45**) (Creer, 1957b, p. 123–124).

The directions of magnetization of the North American Wilberns formation, which was sampled at 10 localities spanning 55 miles, are distributed along a plane passing through the present direction of the earth's field (Howell and Martinez, 1957, p. 390–391). This indicates the presence of varying amounts of a magnetization parallel to the present field, and Howell and Martinez (1957, p. 391) calculate a pole position (**C 47**) from the group of measurements farthest removed from the present field direction. The Sawatch "quartzite" (**C 49**), a sandy dolomite, may have acquired its magnetization during a post-depositional dolomitization, and thus the pole may correspond to the earth's field in post-Cambrian time (Howell and Martinez, 1957, p. 391). A reversal occurs in the Lodore formation (**C 52**) (Collinson and Runcorn, 1960), satisfying the consistency-of-reversals stability test. The Deadwood formation (**C 53**) is cited as Mississippian by Collinson and Runcorn (1960); however, it contains unequivocal late Cambrian fossils at its type locality and is therefore listed here.

Two sets of Cambrian data are available from Australia, one from the Elder Mountain sandstone (**C 55**) and one from the igneous Antrim Plateau basalts (**C 56**) (both from Irving and Green, 1958, p. 66). Details of the sampling and evidence for stability have not been published; concordant results from igneous and sedi-

mentary rocks like these, however, carry much more weight than either result would separately.

Two widely separated poles (**C 29** and **C 31**), closed by stratiform beds, have much more scattered directions. However, most of the directions move toward the other group on

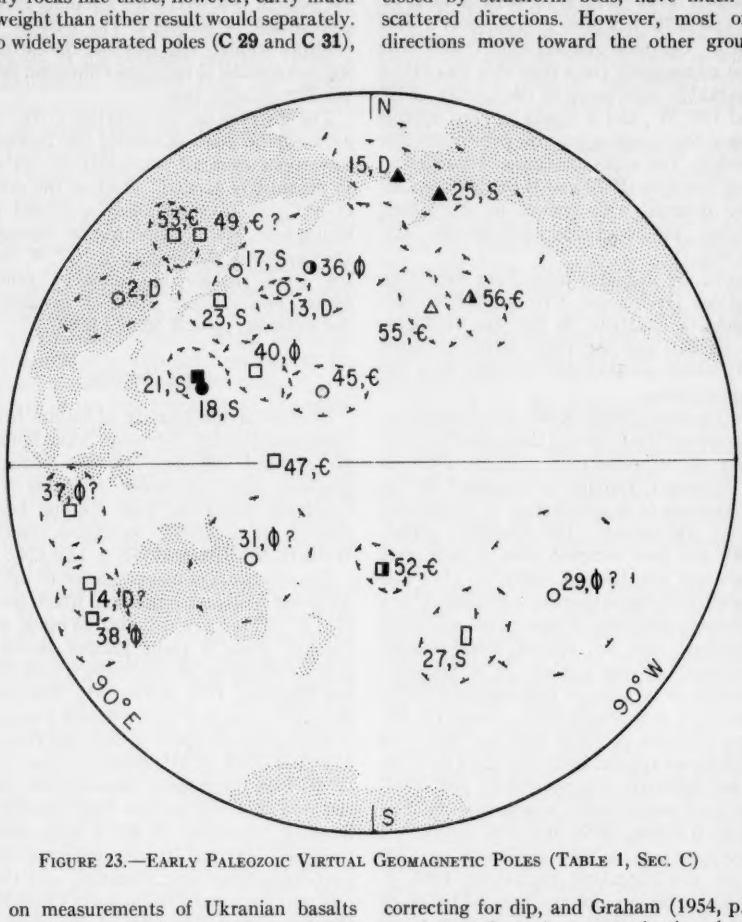


FIGURE 23.—EARLY PALEOZOIC VIRTUAL GEOMAGNETIC POLES (TABLE 1, SEC. C)

based on measurements of Ukranian basalts (Komarov, 1959, p. 1221), are cited as probably Ordovician in age. However, Komarov (1959, p. 1223) states that they may not be the same age, but are both within the lower Paleozoic. Early Ordovician sediments near Leningrad have been sampled by Khramov (1958, p. 185). Pole **C 36** is the mean of a reversed (N) pole and a normal (S) pole that has been "corrected" for an unstable component. Pole position **C 38** (Graham, 1954, p. 219) is based on 45 samples from rocks of the Middle Ordovician Trenton group from New York. Thirty-five of the samples are from undisturbed flat-lying beds and have well-grouped directions of magnetization significantly different from the present field. Ten samples from a distorted zone, encorrecting for dip, and Graham (1954, p. 219) concludes that the magnetization has significant stability. Pole **C 37** is based on directions of magnetization in limestone cobbles of Trenton age in a conglomerate (Graham 1956, p. 738). Since the directions of magnetization are uniform from cobble to cobble, the magnetization must be post-depositional. Agreement with the previous results suggests, however, that not much time elapsed between deposition and magnetization. Pole position **C 40** is based on measurements of red beds in the North American Juniata formation of Late Ordovician age (Collinson and Runcorn, 1960).

The two Silurian units from Europe that have been studied are the Ludlow series (**C 17**) (Creer and others, 1954, p. 165) and the Ural

peridotites (**C 18**) (Komarov, 1959, p. 1222) for which no confidence intervals or other details are available.

Two Middle Silurian formations from North America have been investigated, the Rose Hill formation of Swartz (1923) (**C 21**) (Graham, 1949, p. 148–154) and the Clinton iron ore (**C 23**) (Howell and others, 1958, p. 287–289). The Rose Hill determination is based on 35 oriented samples collected at 6 localities spanning 32 miles. At two localities the samples were collected on the limbs of small folds and at the remaining localities on the limbs of large folds. Directions of magnetization are widely scattered before correcting for dip and, except for three sample directions, are nearly parallel after the dip correction. The deformation occurred near the end of the Paleozoic, and Graham (1949, p. 151) concludes that the magnetization took place before the Permian and very probably at the time of deposition. The Clinton iron ore data are from a sampling area regarded by Howell and others (1958, p. 287) as too small to give a reliable pole position. The rocks have susceptibility anisotropy with the plane of maximum susceptibility nearly in the bedding plane, and the remanent magnetization associated with such a susceptibility anisotropy will probably have a smaller inclination than that of the earth's field in which it developed. The Clinton formation and the Rose Hill formation of Swartz (1923) are both of Middle Silurian age, so that, if the earth's field at this time was that of a dipole consistent with the Rose Hill paleomagnetic results, the field at the Clinton site would have been D = 320, I = −28, or D = 140, I = 28, depending on whether the "normal" or "reversed" sense is taken. The average direction of magnetization of the Clinton rocks is D = 140°, I = 19½°. Since the difference in inclination is in the direction anticipated on the basis of the susceptibility anisotropy, this careful laboratory study by Howell and others supports the results obtained from the study of the Rose Hill formation.

Details of the investigation of the Mugga porphyry from Australia (**C 25**) (Irving and Green, 1958, p. 66) have not been published. The data for the Silurian Red siltstones from China (**C 27**) consist of measurements of specimens from only three oriented samples and, as Chang Wen-You and Nairn (1959, p. 254) state, should be regarded as provisional.

The British Old Red sandstone of Devonian age has been the subject of two independent studies. Clegg and others (1954a, p. 587–588) examined specimens from three oriented samples collected at a single locality (**C 2**) and found that the magnetization was stable in weak D.C. and in strong A.C. magnetic fields. Creer (1957b, p. 113–123) sampled much more widely and noted that flat-lying beds gave consistent results (**C 13**), whereas the directions from folded beds after correcting for tilt were widely scattered, indicating the presence of an unstable component.

The paleomagnetism of one North American formation of Devonian age, the Onondaga limestone, has been studied (**C 14**) (Graham, 1956, p. 738). Directions of magnetization are parallel throughout both layered beds and beds showing penecontemporaneous deformation, indicating post-depositional remanent magnetization. Although the consistency of the results and their divergence from the present field direction point to some stability, the age and origin of the magnetization are uncertain.

The Ainslie volcanic rocks from Australia (**C 15**) (Irving and Green, 1958, p. 66) are probably Devonian but may be Silurian.

In addition, Graham (1954, p. 216) has measured 182 samples from 14 exposures of Ordovician to Permian age in the northeastern United States, with emphasis on the older rocks. Most of these have polarizations within about 30° of the present direction of the earth's field, and there is evidence that some of them are stable and some partially stable. Separate data for each formation are not given (with the exception of that for pole **C 38**), so individual poles could not be computed.

Carboniferous

Many data are available from lava flows, intrusive rocks, sediments, and baked sediments of Carboniferous age from Great Britain. Virtual geomagnetic poles and associated confidence limits for these and other Carboniferous studies are shown in Figure 24. Igneous rocks, probably of late Carboniferous age, and sediments baked by them were sampled at six sites in England spanning 30 miles (**D 5**) (Clegg and others, 1957, p. 220). Samples from adjacent unbaked sediments have random directions of magnetization, whereas the baked sediments have closely parallel directions and intensities of magnetization 100 times larger than those of the unbaked rocks. Pole position **D 8** is from another set of sediments baked by

an intrusive of late Carboniferous age (Clegg and others, 1957, p. 220).

Unbaked sediments from Great Britain that have been studied paleomagnetically include the Gloucester Pennant sandstone (**D 2**) (Clegg and others, 1954a, p. 588), the Pendle Monocline sediments of Lancashire (**D 10**), the Millstone Grit of Lancashire (**D 12**), other Lancashire sediments (**D 11**), and sandstones and siltstones from Derbyshire (**D 16**); the last four are from Belshé's (1957) data. The magnetizations of the 14 specimens from one oriented sample that constituted the data for pole position **D 2** were stable in A.C. fields of several hundred oersted. Stability for the magnetization corresponding to pole position **D 10** is supported by a fold test. The data for the other pole positions (**D 11**, **D 12**, and **D 16**) show considerable scatter, and an additional set of 50 Carboniferous sediment samples from Wales has widely scattered directions and no correlation with the above results (Belshé, 1957, p. 188).

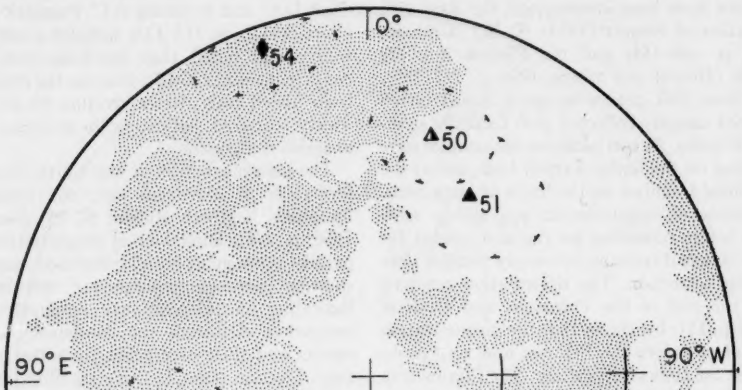


FIGURE 24.—CARBONIFEROUS VIRTUAL GEOMAGNETIC POLES (TABLE 1, SEC. D)

Interbedded with the Derbyshire sediments (**D 16**) are three volcanic units (**D 20**) with well-grouped directions of magnetization which are not significantly different from the direction of the sediment (Belshé, 1957, p. 188). A sequence of 15 lava flows of early Carboniferous age from Scotland falls into three stratigraphic groups, two with approximately parallel magnetizations separated by one with approximately opposed magnetization (Clegg and others, 1957, p. 221). The mean directions of magnetization of these three groups have been used by the present authors to calculate pole position **D 31**.

Samples from two sites 200 yards apart in the Shatterford intrusion (**D 33**), and undoubtedly from the same intrusive unit, have

tightly grouped directions of magnetization almost exactly 180° apart, satisfying the consistency-of-reversals stability test (Clegg and others, 1957, p. 222). Five samples from the Lundy granite, studied by Blundell (1957, p. 191) (**D 36**), are weakly magnetized; however, application of Thellier's test after a month in the laboratory indicates some magnetic stability.

All the paleomagnetic determinations for the Carboniferous of North America are on sedimentary rocks. The Naco formation of Pennsylvanian age was sampled at two localities yielding poles **D 38** (Runcorn, 1956a, p. 309) and **D 40**, (Collinson and Runcorn, 1960). The angular difference between the mean directions of magnetization of these two localities is much less than the difference of either from the present field direction; however, the confidence intervals for these two determinations do not intersect, illustrating the importance of regarding the results from a single sampling site as applying to that site only and not necessarily to the entire formation. The mean directions of magnetization of eight sites spanning 73 miles in the Barnett formation of Mississippian age (**D 43**) are approximately reversed with respect to the mean direction of a ninth site (**D 45**) (Howell and Martinez, 1957, p. 385–388). The magnetization appears to be chemical in origin, and the fair agreement in direction between normal and reversed sites gives evidence of magnetic stability. Pole position **D 47** (Nairn and others, 1959, p. 596) is based on data from the Codroy group of sediments from Newfoundland, and pole **D 49** on combined measurements from the Bonaventure, Kennebacasis, and Bathurst formations of southeastern Canada (Du Bois, 1959b, p. 63).

Two sets of data are available from Australia. Irving (1957b) collected 75 oriented samples from the Kattung varved sediments at four sites (**D 50**). The mean directions of magnetization are very widely scattered before correcting for Permian folding, but are in striking agreement after correcting for dip; moreover, samples from one site are reversely magnetized with respect to those at the other three. Thus both the fold test and consistency-of-reversals test for magnetic stability are satisfied. Pole position **D 51** for the Kattung lavas (Irving and Green, 1958, p. 66) does not differ significantly from that for the varves, and the agreement between the results from igneous rocks and those from sediments also points to stability.

The only results from Africa (**D 52** and **D 54**) (Creer and others, 1948, p. 495) are based on measurements of 19 specimens from four oriented samples which fall into two groups with mean directions 157° apart. In the absence of stability tests, these poles should be viewed with some reservation.

In summary, the virtual geomagnetic poles calculated for the sedimentary formations of Carboniferous age from North America show a remarkable grouping in the vicinity of 36°N., 115°E. The virtual geomagnetic poles calculated from English rocks also fall in this general region but may be subdivided into two groups, one of which agrees very closely with the North American results (poles **D 20, D 2, D 16, D 11, D 10,** and **D 33**) and one of which does not (poles **D 8, D 31, D 5, D 12,** and **D 36**). Two determinations from the latter group (**D 31** and **D 36**) have large ovals of confidence, and the significance of their deviations from the other British group is doubtful. The remaining virtual geomagnetic poles, especially **D 5** and **D 8**, based on intrusive rocks and associated baked sediments, are undoubtedly significantly different. Among several possible interpretations of these two groups of poles, Clegg and others (1957, p. 221–222) considered the following: (a) These rocks may have been magnetized after the Carboniferous; this appears unlikely in view of the evidence relating the magnetization in the baked sediments to the Carboniferous intrusives. (b) The original magnetization may not have been parallel to the earth's field. (c) The earth's field may not have been that of a dipole which remained relatively fixed with respect to the mantle during the Carboniferous. We have tried to fit these data to an axial quadrupole field (*see* Creer and others, 1957, p. 148), for equations of this field) without success. A considerable interval of time is covered by these determinations, however, and it is therefore unlikely that the virtual geomagnetic poles calculated from the North American results, with sampling sites as far apart as Arizona and Newfoundland, would be so well grouped if the earth's field were nondipolar, or if it were changing rapidly during this period. The inconsistency between these two groups of European paleomagnetic poles remains an important unsolved problem.

Viewing the Carboniferous data on a somewhat larger scale, it would be difficult not to recognize that there is excellent evidence for a magnetic field in Carboniferous times different from the present one. Virtual geomagnetic poles calculated from English and American

paleomagnetic directions form a group which is far from random, and the Australian poles, calculated from concordant results from igneous rocks and sediments and with several indications of stability, are certainly significantly different from the North American and European group.

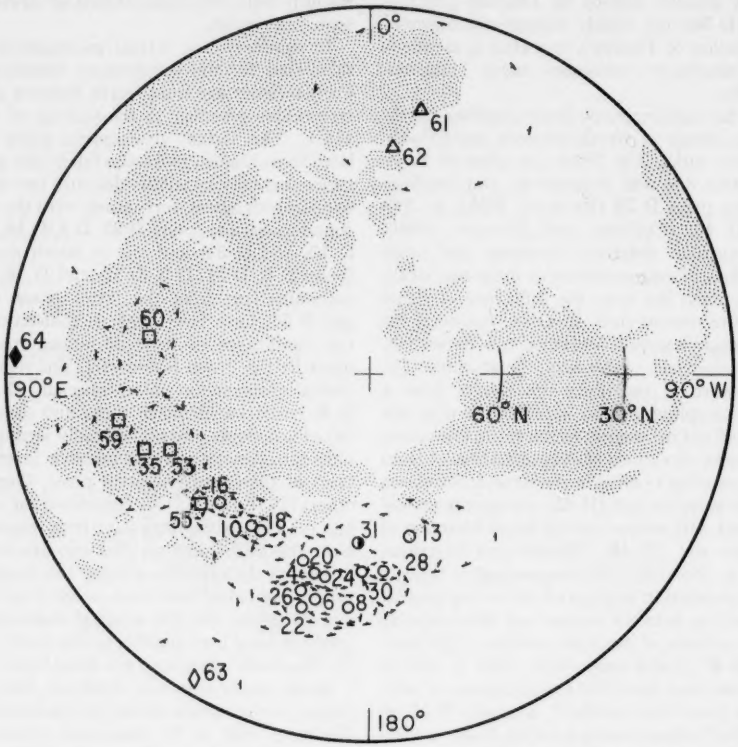


FIGURE 25.—PERMIAN VIRTUAL GEOMAGNETIC POLES (TABLE 1, SEC. E)

Permian

Paleomagnetic results are available from volcanic, intrusive, and sedimentary rocks of Permian age from Great Britain and continental Europe; Figure 25 shows the virtual geomagnetic poles computed from these data. Pole position **E 4** is based on the mean direction of five lava flows from the Exeter volcanic series (Creer, 1957b, p. 112–113). Pole **E 8** is based on data from the Mauchline lavas, and **E 6** on data from the associated Mauchline sedimentary rocks (Du Bois, 1957, p. 178). The Ayrshire kylites (**E 22**) (Armstrong 1957, p. 1277), which intrude the Coal Measures and are cut by Permian volcanic necks, were sampled at five sites.

Poles **E 16**, **E 10**, **E 13**, and **E 18** are based on studies of several units in the Esterel volcanic rocks of southeastern France (Rutten and others, 1957, p. 195; Roche, 1957, p. 2953; As and Zijderveld, 1958, p. 317). **E 16** is based on 14 samples from a single flow and hence a single point in time, and, therefore, the small value of α_{95} does not have the usual paleomagnetic significance. Stability is indicated for the Esterel rocks (pole **E 18**) by the reduction of scatter in directions on partial demagnetization at 150°C. and in an A.C. field of 300 oersted, and also by a fold test.

Two other determinations from European igneous rocks are pole **E 24**, based on measure-

ments of samples from the middle Permian Niedeck porphyry from France (Nairn, 1957b, p. 722), and pole **E 20** based on a number of separate igneous units sampled in the Oslo graben (Rutten and others, 1957b, p. 195).

only reversals in all reported Permian paleomagnetic data—the scatter in the measurements is so extreme as to leave this suggestion unsubstantiated.

The individual virtual geomagnetic poles

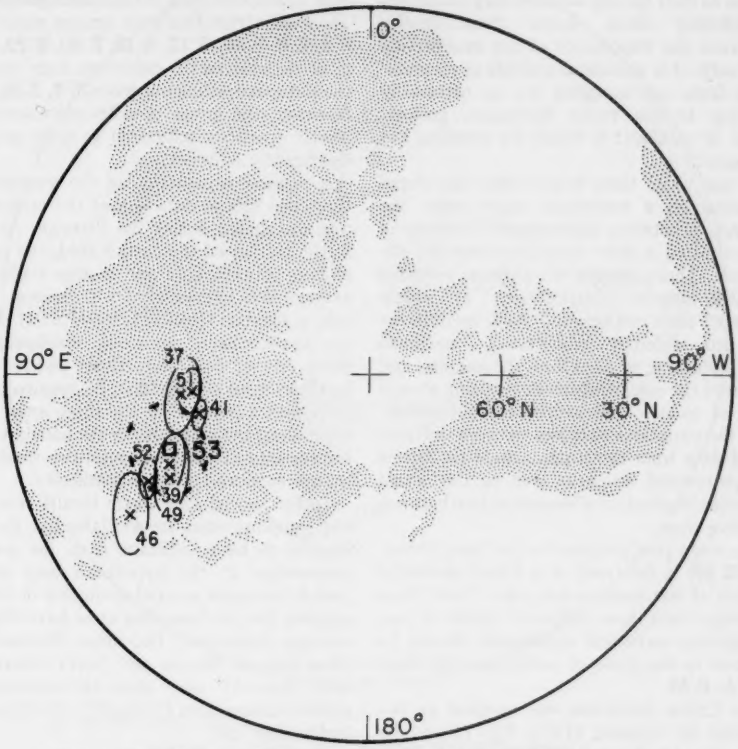


FIGURE 26.—POLES AND CONFIDENCE INTERVALS FOR THE SUPAI FORMATION

Virtual geomagnetic poles corresponding to mean directions at different sampling sites are indicated by X. Numbers refer to entries in section E, Table 1.

Two sedimentary formations from France, the Montcenis and the Saint-Wendel, give poles **E 26** and **E 28**. Twenty-two additional samples from the Montcenis and other sedimentary formations (Nairn, 1957b, p. 722) have directions of magnetization which are either scattered or roughly parallel to the present field; Nairn (1957b, p. 722) regards these as unstable. Late Permian sediments have been sampled at two localities in Central Russia (**E 30** and **E 31**) by Khramov (1958, p. 187). Although reversed magnetizations are suggested for the Tartarskij sediments—the

corresponding to seven separate studies of the North American Supai formation (Permian and Pennsylvanian) at sampling sites spanning several hundred miles are plotted in Figure 26, together with the mean pole position **E 53**. Poles **E 37** (Collinson and Runcorn, 1960) and **E 39** (Graham, 1955, p. 343) correspond to sets of samples collected at about the same locality. Poles **E 46** (Runcorn, 1955b, p. 505) and **E 49** (Doell, 1955b, p. 1167) are based on samples collected at a different locality.

The variations in the pole positions computed from the Supai data give some idea of the

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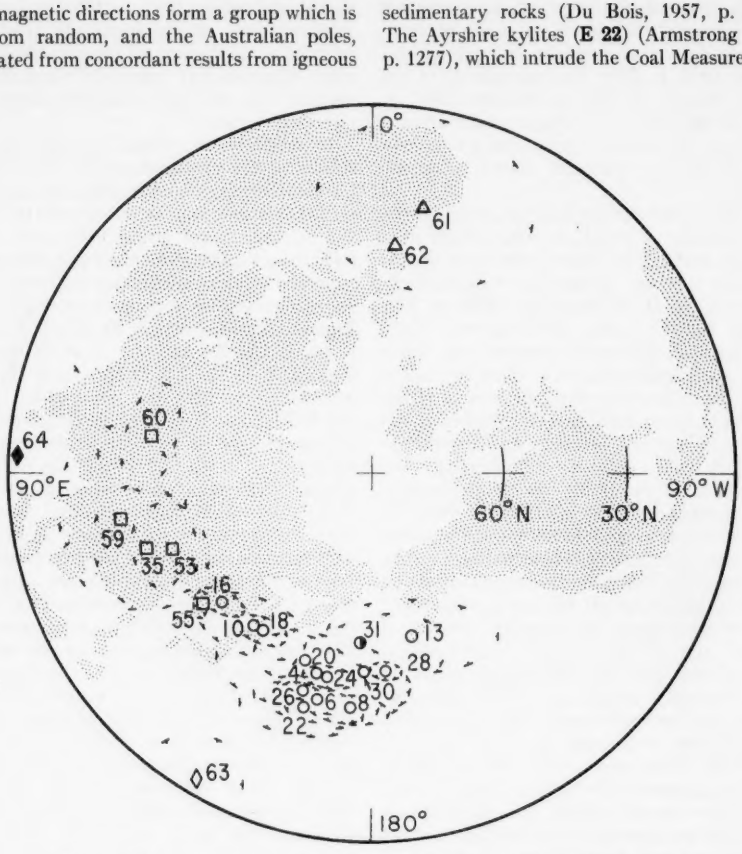

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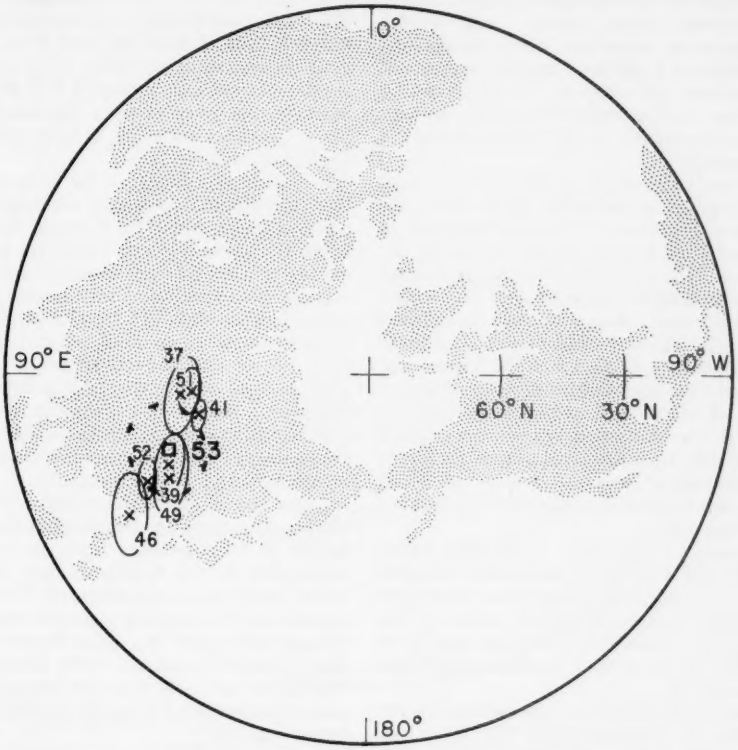

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The variations in the pole positions computed from the Supai data give some idea of the

expectable variations in paleomagnetic studies of sediments and indicate that different populations of magnetic directions have probably been sampled at the different sites. Such differences between two undisplaced points within a formation are entirely normal and possibly are due to their having acquired magnetizations at different times. Again these studies emphasize the importance of not interpreting the results of a statistical analysis of measurements from one sampling site as necessarily applying to the entire formation, geologic period, or continent in which the sampling site is situated.

As one views these results from the Supai formation on a somewhat larger scale, the consistency between the average directions of magnetization is more impressive than the differences. Three groups of workers collected oriented samples independently at widely separated sites, transported them in different ways to different laboratories, machined cubes, cylinders, or discs from the samples, and measured the remanent moments using several types of spinner and astatic magnetometers. Thus the variations observed between different sets of data from the Supai include the effects of experimental errors as well as the actual variations in direction of magnetization between sampling sites.

The mean pole position for the Supai formation (**E 53**) is the result of a Fisher statistical analysis of the seven mean poles. Since these individual data have different values of α_{95}, no rigorous statistical significance should be attached to the circle of confidence calculated for pole **E 53**.

The Cutler formation was sampled at two localities by Graham (1955, Fig. 7) and by Collinson and Runcorn (1960). Pole **E 35** is midway between the two mean poles for these sets of data; "α_{95}" is taken as half the angular distance between the two poles but has no other significance. The Abo formation (**E 59**) is similarly based on data from two sites, and the Yeso formation (**E 55**) and the Sangre de Cristo formation (**E 60**) are each based on samples from a single site (all calculated from data of Graham, 1955, Fig. 7). The Sangre de Cristo formation includes rocks of middle or late Pennsylvanian age as well as early Permian age.

The two Australian virtual geomagnetic poles are based on samples from three lava flows from the Upper Marine volcanic series (**E 61**) and one flow from the Lower Marine volcanic series (**E 62**) (Irving and Green, 1958, p. 66). The two widely spaced poles from Africa are based on sediments of late Permian (**E 63**) and early Permian (**E 64**) age (Creer and others, 1958, p. 495).

A striking feature of the Permian paleomagnetic results is the separation of European from North American virtual geomagnetic poles. The results from European igneous rocks (poles **E 4, E 8, E 10, E 13, E 18, E 20, E 22, E 24**) show no systematic difference from those of the European sediments (poles **E 6, E 26, E 28**) However, the group of poles calculated from North American sediments is quite probably significantly different.

A possible explanation of the separation of these two sets of data is that the earth's field was not dipolar during the Permian. As noted in the section on the earth's field, the relative motion between core and mantle leads to an average field symmetrical with respect to the axis of rotation (Runcorn, 1959, p. 91). To test the hypothesis of an axial, nondipole field, three sets of data from three widely spaced localities would, in general, be required; intersection at a point P of the three great circles lying along the directions of the horizontal field components at the three localities would constitute support for the hypothesis.

In the special case where results from only two sampling areas are available but the areas happen to be equidistant from the point of intersection P, the hypothesis may also be tested, inasmuch as axial symmetry of the field requires that the sampling areas have the same average inclination. The mean Permian sampling areas in Europe and North America are both about 62° away from the corresponding point of intersection P (Fig. 27). The two sets of inclinations are:

Europe:	−9°, −6°, −4°, −16°, −13°, −22½°, −16°, −33°, +1½°, −7°, +6°, −9°
North America:	+6°, +32½°, +10°, +9½°, +2°, +23°, +8°, +5°, +18°, −1°, +8°, +55°, +30½°

These differences in inclination between Europe and North America appear to be systematic, and therefore an axially symmetrical nondipole field does not appear to be a suitable explanation for the differences between European and North American virtual geomagnetic poles.

A second striking feature of the North American and European Permian results is that all the field directions have the same polarity—

there are no "reversals." (The Tartarskij sediments may be an exception.) If the field-reversal hypothesis is incorrect, the interpretation that must be made is that mineral assemblages necessary for self-reversal are abundant in Carboniferous and Triassic rocks since both these periods have many reversals, but are missing in all (or almost all) Permian rocks studied to date. The interpretation of this feature on the basis of the field-reversal hypothesis is that oscillations in the polarity of the earth's field have been intermittent. The latter interpretation appears to us somewhat more plausible.

The two results from the Permian of Australia are consistent with each other but not with the general grouping of results from North America and Europe. Although these Australian results are based on estimates of the field at only four points in time during the Permian, they are consistent with the excellent Australian Carboniferous data.

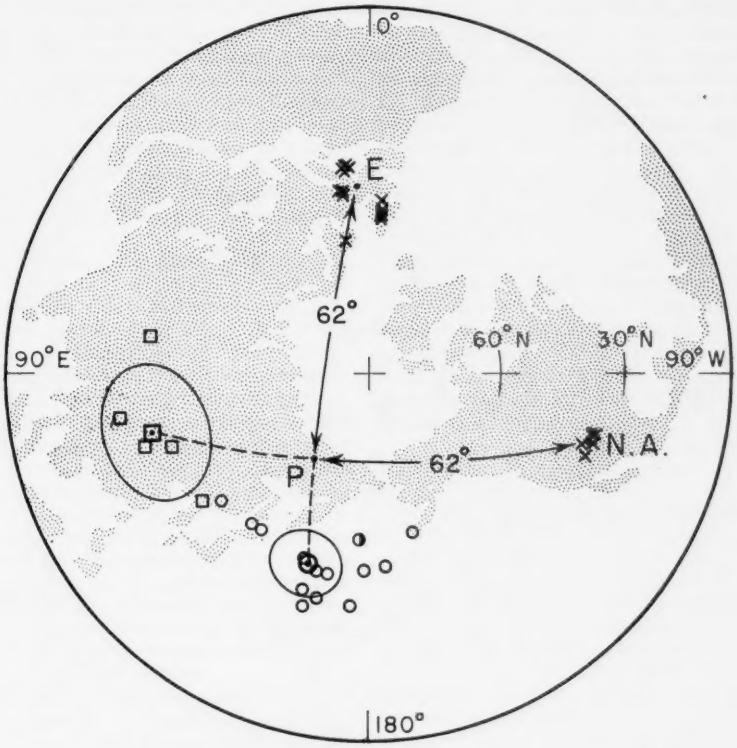

FIGURE 27.—TEST FOR AXIAL, NONDIPOLE FIELD

P is the intersection of great circles connecting mean sampling sites and mean virtual geomagnetic poles, as indicated by a heavy square (North America) and a heavy circle (Europe).

Triassic

The Triassic poles are plotted in Figure 28. The nine sampling sites in the Keuper Marl sandstones from England (Clegg and others, 1954a; 1954b, p. 195) fall into two groups which have directions of magnetization approximately reversed with respect to each other. These sediments are stable in weak D.C. fields and in strong A.C. fields and satisfy a fold test for rather shallow dips. Pole **F 12** is based on

the nine mean site directions as listed in the original paper. Creer (1957a, p. 136) showed that Keuper Marl samples collected at many other localities have two components of magnetization, one stable, the other unstable; the stable one is essentially parallel to that described by Clegg.

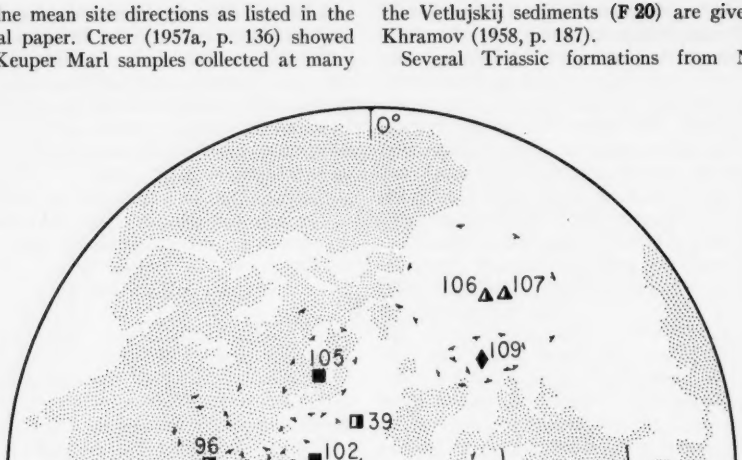

FIGURE 28.—TRIASSIC VIRTUAL GEOMAGNETIC POLES (TABLE 1, SEC. F)*

Pole **F 13** combines the results of measurements on sediments of early Triassic age from seven sites in the Vosges region of France. There is considerable scatter in directions of magnetization, which has been interpreted by Clegg and others (1957, p. 225) to be indicative of partial instability. Of the seven localities in Spain where sedimentary samples were collected, only three localities had consistent directions of magnetization (**F 15** and **F 16**) (Clegg and others, 1957, p. 225–226). Pole (**F 16**) combines data from two sites. Data for the Vetlujskij sediments (**F 20**) are given by Khramov (1958, p. 187).

Several Triassic formations from North America have been sampled extensively at a large number of sites, and in order to present all the data on a single plot only one mean pole position is shown for each formation; for intraformational details the reader is referred to Section F of Table 1. Pole **F 39** is based on work by several workers at eight sampling sites in the Chinle formation or formations of equivalent Late Triassic age (Graham, 1955, Fig. 7; Collinson and Runcorn, 1960). At some of the sites an unstable component of magnetization is probably present (Collinson and Runcorn, 1960), and our use of the actual measured directions in the statistical analysis rather than the direction of an inferred stable component may have resulted in a position too far north for pole **F 39**. Five of the sites are normally mag-

netized, one site is probably reversed, and at two sites most of the samples are normal, but several are reversed.

Pole **F 79** is based on a very extensive sampling of the Chugwater formation by Collinson and Runcorn (1960). At 7 of the 10 widely spaced sampling sites both normal and reversed groups of magnetizations were encountered; in the statistical calculations for pole **F 79** each group, whether normal or reversed, was treated as a separate measurement. The scatter of poles corresponding to the mean site directions is unusually small (k = 57), and these data show excellent stability by the consistency-of-reversals test.

The Moenkopi formation of Early and Middle (?) Triassic age (**F 91**) has been sampled at seven sites by Runcorn (1956a, p. 311–312), Kintzinger (1957, p. 931), and Collinson and Runcorn (1960). The four normally magnetized and the three reversely magnetized sites appear to form two separate groups; the poles of the normal group are north of the others. Thus the consistency-of-reversals test is not satisfied, indicating the presence of a comparatively small unstable component of magnetization. However, as pointed out by Creer and others (1957, p. 151), the average position of the two groups probably corresponds reasonably well to the stable component of magnetization. The Springdale sandstone member of the Moenave formation (**F 22**) (Runcorn, 1956a, p. 312–314) was sampled at one site and is probably partially unstable.

Poles **F 93** and **F 96** are based respectively on the magnetizations of lava flows and on those of sediments and lava flows combined, all from the Connecticut Valley (Du Bois and others, 1957, p. 1186). Pole **F 98** (Du Bois and others, 1957, p. 1186) is based on data from sediments at the top of the Newark group, and pole **F 100** (Graham, 1955, Fig. 7) on sediments from the bottom of this same group. Lavas from Nova Scotia were sampled at 12 sites by Bowker (letter of November 16, 1959, to Doell), and pole **F 102** is the mean of the virtual geomagnetic poles calculated for each of the 12 sites. The Dinosaur Canyon sandstone member of the Moenave formation, studied by Kintzinger (1957, p. 931) (**F 105**), is Late Triassic(?).

Pole **F 106** was calculated from measurements of the average inclination of unoriented vertical core samples from Tasmanian volcanic tuffs (Almond and others, 1956, p. 775); the steep inclination of 81½° implies a virtual geomagnetic pole within 16° of the sampling site. Deviations of the core hole from the vertical, a common occurrence, may contribute an additional uncertainty. Pole **F 107** (Irving and Green, 1958, p. 66) is based on measurements of oriented surface samples of the Brisbane tuff, and it is of interest to note that the pole inferred from the unoriented vertical core samples is in agreement with this pole. Pole **F 109** is based on sediments of Late Triassic age from Africa (Nairn, 1957a, p. 166–167); there is good internal consistency in this study, but contemporaneous sediments from adjacent localities have widely scattered directions.

All but one of the poles calculated from Triassic paleomagnetic measurements are significantly displaced from the present geographic pole. The North American poles are very roughly grouped at about 60° N. and 105° E., but many poles for individual formations are displaced considerably from this position. Reversals in both igneous and sedimentary formations are common.

Jurassic

Jurassic virtual geomagnetic poles are plotted in Figure 29. Pole **G 2** is based on results from English rocks which may be unstable (Nairn, 1957c, p. 311–312). Pole **G 5** is based on measurements of six oriented samples from several British Lower Jurassic formations; five of the six samples had reversed polarity (data are from Nairn, 1957a, p. 311–312). Poles **G 6** and **G 7** are based on two groups of sedimentary samples from the Early Jurassic (upper Lias) in Britain (Girdler, 1960, p. 358–359). One group is very nearly reversed with respect to the other, indicating stability by the consistency-of-reversals test. Samples from other nearby sites were unstable (Girdler, 1960, p. 354–355). Measurements on volcanics from the Pyrenees gave pole **G 8** (Girdler, 1960, p. 359–361). The flows are dated as Early Jurassic on the basis of intercalated limestone beds. Poles **G 10** and **G 12** are based, respectively, on measurements of radiolarite and limestone samples of Middle Jurassic age from the Alps (Hargraves and Fischer, 1959, p. 38).

From North America the Kayenta formation has been sampled at five sites (Collinson and Runcorn, 1960), and pole **G 20** is the mean of the virtual geomagnetic poles for each site. The Carmel formation (**G 21**) (Collinson and Runcorn, 1960) was sampled at one locality.

The Karroo dolerites were sampled exten-

sively both in tunnels and mines (**G 28**) and on the surface (**G 31**) (Graham and Hales, 1957, p. 155). The surface samples have more scattered directions of magnetization, but the mean directions for both groups agree. Some of the underground sills are magnetized reversely with respect to the others, and associated baked sediments are invariably magnetized in the same direction as the sill that baked them. Such a relationship points strongly to reversal of the earth's field. Pole **G 36** for the Karroo basalts (Nairn, 1956, p. 936; 1957a, p. 166), which are as old as the dolerites or slightly older, does not differ significantly from the pole for the dolerites.

Basalts from the Parana basin of South America, as well as sandstones baked by them, have been studied by Creer (1958, p. 377). The two sets of results agree with each other and have been combined in calculating pole **G 52**. The confidence interval of 3° by 1½° (Creer, 1958, p. 385) is probably too low, while the one shown in Figure 29 of 16° by 9½°, estimated by us, may be too large. The correct value depends on how many independent points in time (*i.e.*, separate lava flows) were sampled.

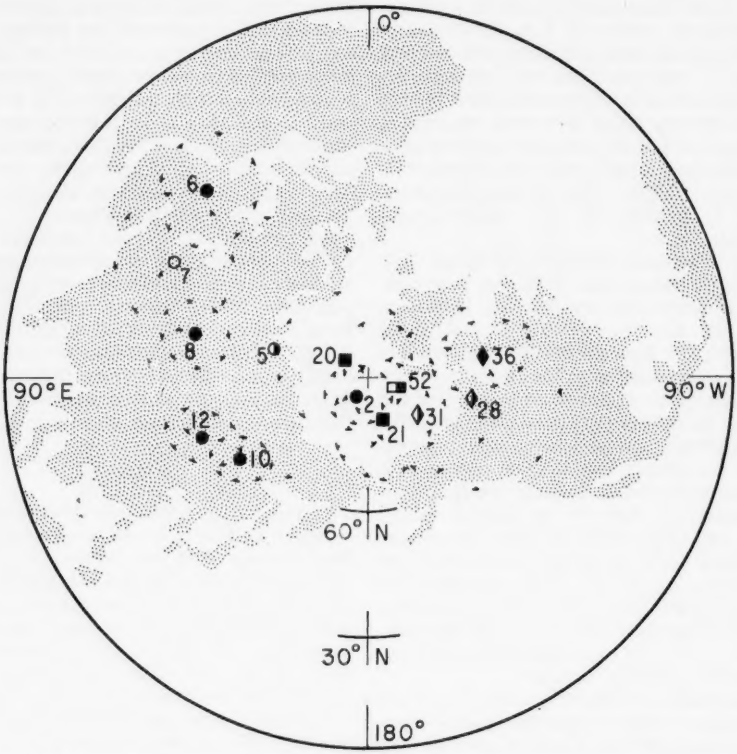

FIGURE 29.—JURASSIC VIRTUAL GEOMAGNETIC POLES (TABLE 1, SEC. G)
Poles H 5, H 7, and H 10 (shown in Fig. 30) may also be Jurassic and possibly should be included here.

In addition to the points plotted in Figure 29, the Rajmahal traps of India (pole **H 10** in Fig. 30) and the Tasmanian dolerites (poles **H5** and **H 7** in Fig. 30) may also represent Jurassic field directions—the ages of these rocks are uncertain.

In addition to the Jurassic measurements reported above, four samples from the Summerville formation and three from the Carmel formation were measured by Torreson and others (1949, p. 125) in Utah. Because of the extreme scatter in the measured directions,

no poles could be computed. The Jurassic virtual geomagnetic poles calculated for all continents show considerable scatter, and a nearly parallel to the present field. Since no numerical data were given a pole could not be calculated.

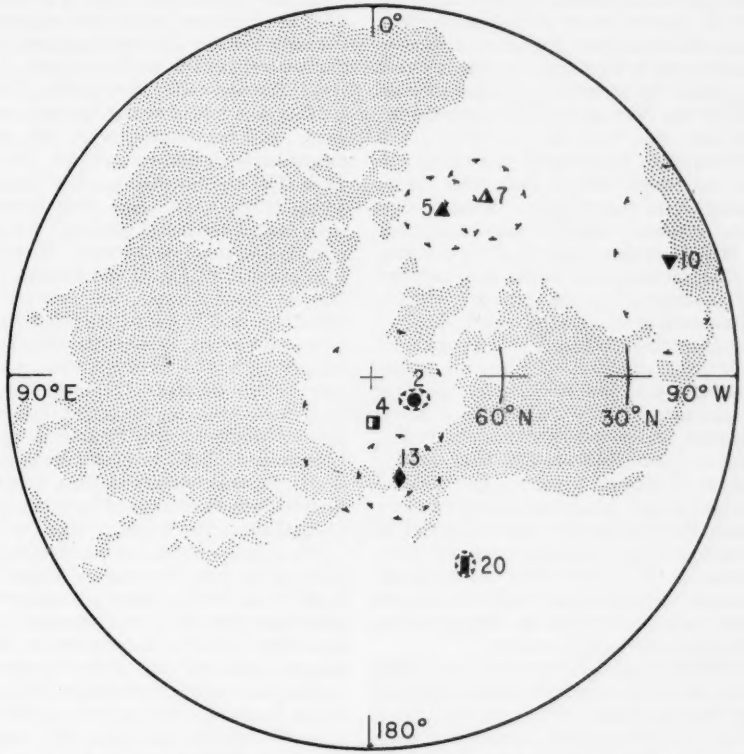

FIGURE 30.—CRETACEOUS VIRTUAL GEOMAGNETIC POLES (TABLE 1, SEC. H)
Poles H 5, H 7, and H 10 may be Jurassic

relatively large number are not significantly different from the present geographic pole.

Cretaceous

Only one Cretaceous pole from Europe is available, that from the Wealden sediments of the Isle of Wright (**H 2**) (Wilson, 1959, p. 753). Several lines of evidence point to the existence of a stable component of magnetization about 4° from the present field direction. Nairn (1957c, p. 309) reports that the remanent magnetization of a red band 2–6 inches thick at the base of the Lower Cretaceous Gault Clay formation from England has both reversed and normal directions of magnetization which are

The only Cretaceous paleomagnetic data from North America for which a pole can be calculated consist of measurements made on 6 oriented samples from the Dakota sandstone from the Colorado Plateau (Runcorn, 1956a, p. 314–315). Three of the samples have magnetizations approximately reversed with respect to the direction of the present field, and 10 specimen measurements from these samples are the basis for the generally cited North American Cretaceous pole. For pole **H 4** (Fig. 30) we have used the mean sample directions for the three reversed and two normal samples; one obliquely magnetized sample has been excluded.

The Tasmanian dolerite sills cut Upper Trias-

sic (or possibly Jurassic) rocks and are displaced by early Tertiary faults. Although they are thus no more closely dated than Late Triassic to early Eocene, the paleomagnetic results are usually plotted as Jurassic. We have taken the liberty of placing these virtual geomagnetic poles on the Cretaceous diagram to emphasize the uncertainty in their ages. The dolerites have been extensively sampled at surface outcrops (**H 5**) (Irving, 1956b, p. 166–167), and measurements have also been made on vertical well cores unoriented with respect to azimuth (Almond and others, 1956, p. 773). Since the inclination of the magnetization in the cores is $-85\frac{1}{2}°$, it follows that the virtual geomagnetic pole **H 7** lies within about 9° of the sampling site. This interpretation of the core measurements is supported by the agreement with the measurements of the oriented surface samples.

The lower Rajmahal traps of India are interbedded with sediments of Early to Middle Jurassic age (Krishnan, 1956, p. 273); the upper traps are petrologically similar to post Lower Cretaceous Deccan traps to the west (Hobson, quoted in Pascoe, 1929, p. 146) but otherwise appear to be undated. In paleomagnetic studies these results have usually been assigned to the Jurassic. However, we have plotted pole **H 10** for this formation as calculated from the measurements to Clegg and others (1958), on the Cretaceous diagram, again with a view to preventing data of uncertain age from becoming fixed in the paleomagnetic record.

Pole **H 13** is based on extensive investigations of lavas and dikes from Madagascar belonging to the Turonian stage of the Cretaceous (Roche and Cattala, 1959, p. 1050). Nagata and others (1959, p. 381) have calculated pole **H 20** from measurements on the Inkstone red shales of Middle to Early Cretaceous age from Japan.

Eocene

Most of the early Tertiary formations studied paleomagnetically are volcanic rocks that are imprecisely dated: an unfortunate circumstance, since paleomagnetic interpretation of these results depends on good stratigraphic control. Listed in Section I of Table 1 and shown in Figure 31 are all Tertiary results which have been described as Eocene, possibly Eocene, or early Tertiary, even though the balance of the evidence in some instances may favor assignment to a post-Eocene age. This procedure has undoubtedly resulted in the inclusion of some post-Eocene data in Section I and Figure 31, and therefore some of these virtual geomagnetic poles are expected to be near the present geographic pole, in agreement with the other post-Eocene results.

The age of the Lundy dikes (**I 2**) (Blundell, 1957, p. 191) is not known, but assignment to the early Tertiary has been suggested because the magnetization is similar to that of other lower Tertiary rocks from Britain. Thellier's stability test, with 1 month between measurements, and the consistency of this reversed magnetization point to stability. The Arran dikes (**I 3**) are described as early Tertiary by Irving (1959, p. 64). The Mull lavas (**I 7**) have been described as probably Eocene but possibly Oligocene or Miocene (Hospers and Charlesworth, 1954, p. 41–42; Hospers, 1955, p. 71); only flows showing some evidence of stability were used in calculating pole **I 7** (Bruckshaw and Vincent, 1954, p. 584–585). The Mull intrusive rocks (**I 4**) (Vincenz, 1954, p. 593) are younger than the lavas. The Antrim basalts are described as probably Eocene or Oligocene but possibly Miocene or early Pliocene (Hospers and Charlesworth, 1954, p. 41–42; Hospers, 1955, p. 71). Separate studies of these basalts have been made by Hospers and Charlesworth (1954, p. 40) (**I 10**) and by Wilson (1959, p. 752) (**I 20**).

The early Eocene age of the Siletz River volcanic series from the western United States (**I 22**) (Cox, 1957) is based on extensive fossil collections from sediments intercalated with the lava flows. Stability is indicated by the consistency-of-reversals test and by a reduction of scatter on correcting for folding which is mostly of late Eocene age. Irving (1959, p. 59) has calculated a pole for the Laney shale member of the Green River formation based on data of Torreson and others (1949, p. 125). We have recomputed this pole (**I 24**), and our results agree essentially with Irving's; in addition we have computed two other poles, **I 25** and **I 26**, from measurements on other sediments of Eocene age reported by Torreson and others (1949, p. 125). These three flat-lying units have magnetizations nearly parallel to the present field, with no reversals.

Pole **I 27** is based on an extensive sampling of the older volcanics of Victoria (Australia) which are of early Tertiary and probably Eocene age (Irving and Green, 1957, p. 351). The Tasmanian basalts have been described as Oligocene or Miocene by Almond and others (1956, p. 771), but Banks (1958, personal communication) believes they may be Eocene; vertical cores from a bore hole have an average inclination of $-83°$,

indicating a virtual geomagnetic pole (**I 28**) within 13½° of the sampling site.

The Deccan traps of India have been studied independently by several workers. Pole **I 30** is based on seven oriented samples collected from different levels over a wide area (Irving, 1956a, p. 40). Clegg and others (1957, p. 227–230) and Deutsch and others (1959) have investigated the traps in great detail and have collected from several levels at seven sites covering a wide area. Results from horizons which are topographically and probably stratigraphically nearer the bottom of the sequence have been combined for pole **I 59**, and those from the upper horizons for pole **I 58**. Results from a post Deccan trap tuff form the basis for pole **I 57**.*

In addition to the Eocene measurements reported above, five samples from the Wasatch formation in Colorado and four from Wyoming have been measured by Torreson and others (1949, p. 125). The nine measurements were much too scattered to permit the calculation of a pole.

In summary, the virtual geomagnetic poles calculated from the magnetizations of rocks of Eocene (and probably younger) age fall into two groups, one near the present geographic pole and the other distributed north of lat 30° N. and within about 25° of the 300th meridian. Three poles in the first group correspond to sedimentary formations from North America having no evidence for stability, and the remaining poles in this group correspond to volcanic rocks of somewhat uncertain age from Europe.

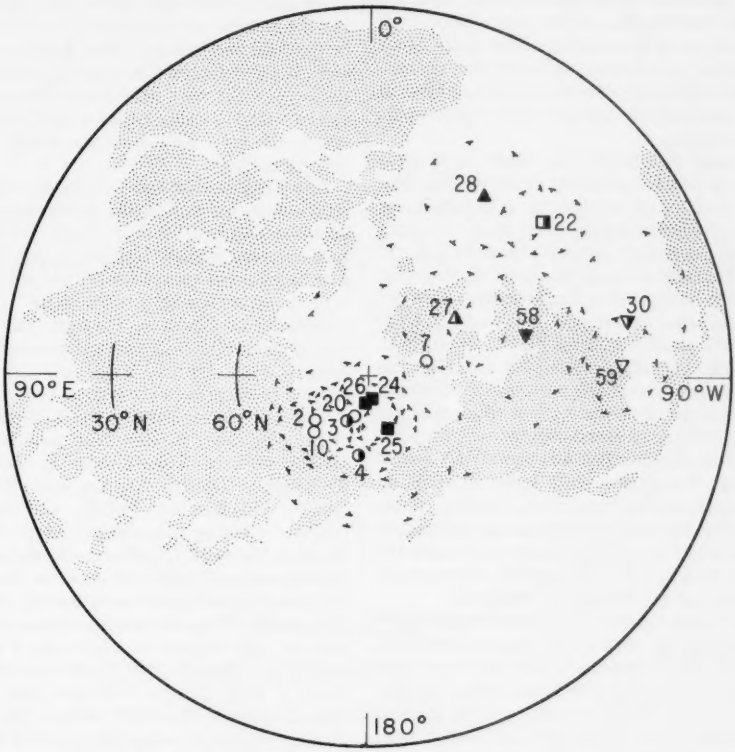

FIGURE 31.—EOCENE AND EOCENE (?)VIRTUAL GEOMAGNETIC POLES (TABLE 1, SEC. I)
Poles for all paleomagnetic studies of rocks described as Eocene, possibly Eocene, or early Tertiary are included. Some are almost certainly post-Eocene.

* Pole **I 57** was inadvertently not plotted.

Polar Wandering and Continental Drift

General statement.—Paleomagnetic measurements, which have been discussed in terms of their equivalent virtual geomagnetic poles, tell us only the direction of the earth's field at the time and place of formation of the rocks. If we are to apply these results to the problems of continental drift and polar wandering we must in some way relate the magnetic field to geographic configurations.

In order to interpret the results from some period in terms of continental drift we must first have enough data to know the configuration of the magnetic field during that period: was it, for example, essentially a dipole field as at present? It is further necessary to assess the expected variation between sampling areas that have not been displaced with respect to each other. Only then is it possible to infer that a departure in direction at some given sampling area indicates displacement of that area with respect to others. Polar-wandering interpretations require further that there be some connection between the magnetic-field configuration and the earth's axis of rotation. Continental-drift interpretations do not require such a connection, but they do require that the field configuration be changing at a rate which is small in comparison with the degree to which we can establish the contemporaneity of the formations compared.

Axial dipole theory.—The paleomagnetic evidence indicates, with a high degree of precision, that the earth's magnetic field, when averaged over a few thousand years, was dipolar in nature and parallel to the present axis of rotation throughout early Pleistocene and Recent time. It follows that the two axes were parallel during this time if we can establish that the earth's axis of rotation has also remained fixed. The distribution of the Pleistocene polar ice sheets defines rough limits for possible movements of the rotational axis during this time, but it is doubtful whether polar wandering of less than 10°–20° could be detected. Estimates of displacements of the rotational axis during the past several decades, as found from astronomical observations, indicate (Elsasser and Munk, 1958, p. 230) that the geographic pole moved at most 15 feet between 1900 and 1940. If the average rate of motion during the past half million years had been twice that amount, the total polar shift would have been only 1°. Since there is no other evidence to suggest that the axis of rotation differed significantly from the present one, the late Pleistocene and Recent paleomagnetic results constitute strong evidence in support of the dynamo theory for the origin of the earth's magnetic field.

It is, of course, possible that the coincidence of the paleomagnetically determined dipole axis and the present rotational axis is fortuitous. The earth's field may not have been dipolar or axial, and displacements of the sampling areas or of the entire crust may have been of exactly the right amount to compensate for such irregularities in the field. This, however, would be extremely unlikely.

Late Tertiary polar wandering and continental drift.—As noted earlier, the paleomagnetic results for Oligocene through early Pleistocene time are very similar to those for late Pleistocene and Recent except for the presence of reversals and somewhat greater scatter in mean pole positions. However, it is much more difficult to determine on geological or geophysical grounds what relationship the axis of rotation has had to the present continental configuration during this same interval. Large displacements of the pole have, in fact, been postulated by several workers. Polar-wandering curves of Kreichgauer (1902), Köppen and Wegener (1924), Milankovitch (1938), and Köppen (1940) as given by Gutenberg (1951, p. 202) are shown in Figure 32. (Note that only the paths of Kreichgauer and Milankovitch are in relation to the present continental configuration; those of Köppen and Wegener assume continental drift and are with respect to Africa only.) On the other hand, Chaney (1940, p. 486) and Durham (1950, p. 1260; 1952, p. 339) have cited paleoclimatological evidence to indicate that early Tertiary isoclimatic zones were parallel to present latitude lines. Further clarification of the paleobiogeographic picture during this interval is greatly to be desired. To the extent that we are willing to extrapolate the axial dipole model back into the past, the paleomagnetic evidence indicates that no large shift of the axis of rotation has occurred during the late Tertiary—a conclusion previously arrived at by Hospers (1955, p. 72–73) from analysis of fewer data.

Some recent theories calling for substantial shifts of the pole of rotation in late Tertiary time have cited paleomagnetic evidence in support of such shifts (Ewing and Donn, 1956, p. 1065; Hapgood, 1958, p. 308). Although some Oligocene to early Pleistocene virtual geomagnetic poles do, in fact, have ovals of confidence that do not include the present pole, there are several reasons why these probably do not indicate polar wandering or displacements of the

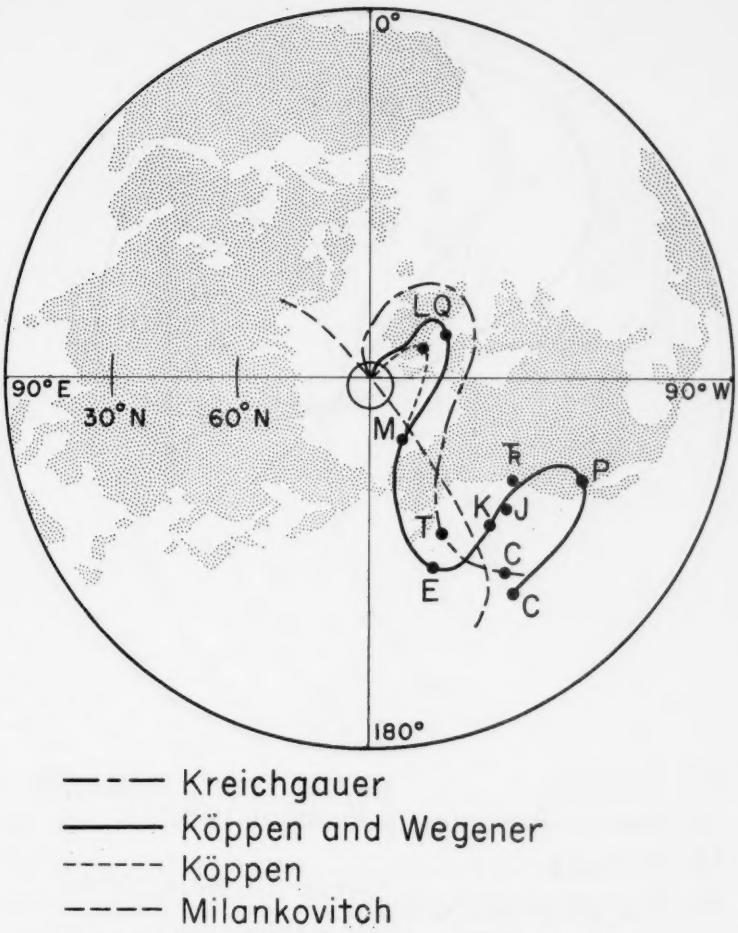

FIGURE 32.—POSTULATED TERTIARY POLAR-WANDERING PATHS
C—Carboniferous, P—Permian, T—Triassic, J—Jurassic, K—Cretaceous, E—Eocene, M—Miocene, LQ—lower Quaternary. (After Gutenberg, 1951)

sampling areas. For statistical reasons previously discussed, some of the confidence intervals may be too small. Pole **A 82**, for example, is based on data with widely varying confidence intervals, and a circle of confidence cannot be rigorously calculated from such data. Moreover, the virtual geomagnetic poles that are displaced from the present geographic pole are distributed throughout the time interval represented and are not confined to older rocks. This suggests that the effect may be random rather than systematic. One also expects, at the 95 per cent probability level used in these analyses, that 1 out of every 20 ovals of confidence will appear to be significantly different. Thus, although small amounts of polar wandering cannot be excluded, the paleomagnetic evidence appears to us to offer no support for theories requiring substantial late Tertiary polar wandering; on the contrary, it indicates that the pole has remained relatively fixed during this time.

Paleomagnetic polar wandering and continental drift.—Since the average magnetic dipole and rotational axes have been parallel in late Pleisto-

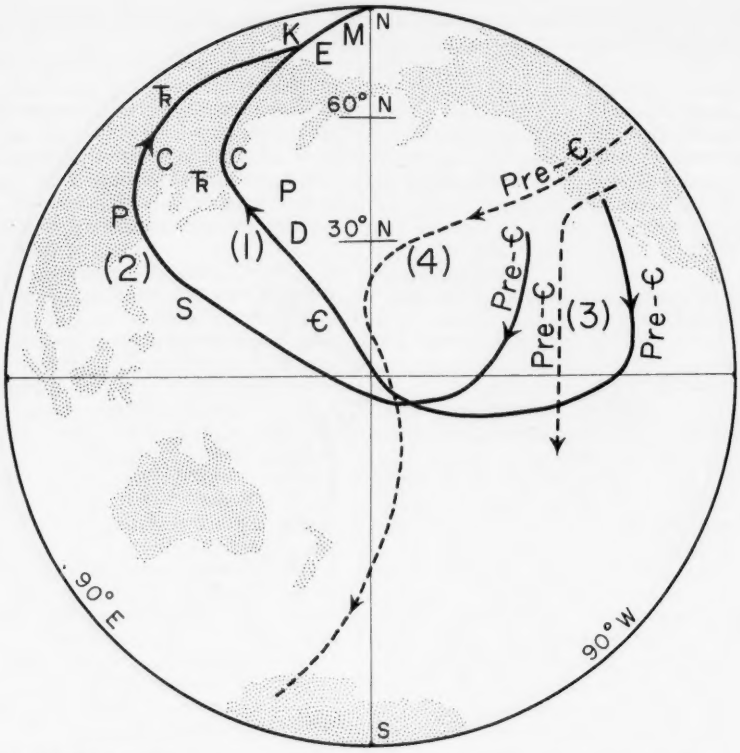

FIGURE 33.—POSTULATED PALEOMAGNETIC POLAR-WANDERING CURVES FOR EUROPE AND NORTH AMERICA
Pre-€—Precambrian, €—Cambrian, S—Silurian, D—Devonian, C—Carboniferous, P—Permian, T—Triassic, K—Cretaceous, E—Eocene, M—Miocene

cene through Recent time, and probably since the Oligocene, the principle of uniformitarianism suggests polar wandering and continental drift as possible interpretations for paleomagnetic results that do not agree with the present field configuration. Such an interpretation, in the form of a polar-wandering curve, was suggested by Creer and others (1954, p. 165) to explain pre-Tertiary paleomagnetic data from North America and Europe. When more data became available, the possibility of obtaining a better fit with separate paths for North America and Europe was pointed out by Irving (1956a, p. 39) and Runcorn (1956b, p. 82–83); these curves are shown in Figure 33 (after Creer and others, 1957, p. 147). The more westerly path inferred from the North American data was explained as due to a drift of North America of some 24° away from Europe prior to the middle Tertiary.

Du Bois has traced Precambrian paths of polar wandering for North America and Europe which cross the other paths nearly at right angles (Fig. 33); he interprets these and the later

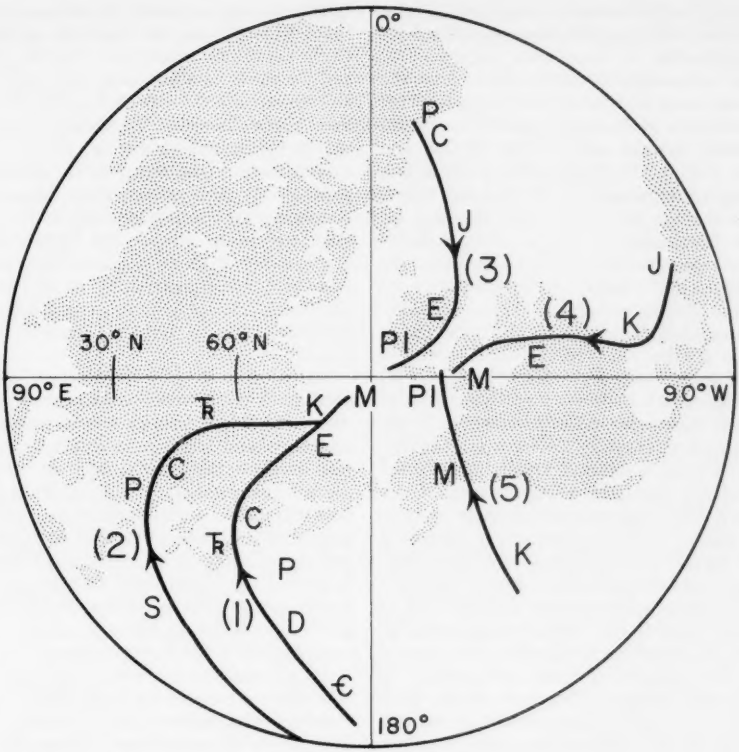

FIGURE 34.—POSTULATED PALEOMAGNETIC POLAR-WANDERING CURVES FOR EUROPE, NORTH AMERICA, AUSTRALIA, INDIA, AND JAPAN

€—Cambrian, S—Silurian, D—Devonian, C—Carboniferous, P—Permian, T—Triassic, J—Jurassic, K—Cretaceous, E—Eocene, M—Miocene, Pl—Pliocene

results as due to a westward drift of North America with respect to Europe of 45° (Du Bois, 1957, p. 179; 1958, p. 512).

Subsequent paleomagnetic data from India, Australia, and Japan have led many authors to postulate different polar-wandering curves for each of these regions. The Paleozoic and later portions of these curves are reproduced in Figure 34, and one may readily note that very large relative drifts and rotations are required to bring them into coincidence. The Indian studies, mostly on the Deccan traps, have been interpreted by a number of workers as indicating that India has rotated about 24° counterclockwise and has drifted 4000–5000 km with respect to North America and Europe since the

Eocene (Clegg and others, 1956, p. 430; Irving and Green, 1957, p. 358; Deutsch and others, 1959, p. 53–54).

The paleomagnetic results from Australia have also been interpreted as evidence for large displacements of Australia, again with respect to North America and Europe (Irving and Green, 1958, p. 71; Irving, 1959, p. 69–72). A suggested interpretation for the Japanese results is one of polar wandering along the path for North America and Europe, on which is superimposed the effects of a drift of Japan since the Cretaceous and a fairly large rotation since the Miocene (Nagata and others, 1959, p. 382–383).

Paleomagnetic evidence has also been cited in support of other rotations and displacements of land masses. Nairn and others (1959, p. 596) have suggested a 20-degree counterclockwise rotation of Newfoundland with respect to North America on the basis of a comparison between sets of Carboniferous and Precambrian virtual geomagnetic poles from these two areas. Creer (1958, p. 389) suggests a drift of South America with respect to Africa from a comparison of some Jurassic measurements from these continents, and Creer and others (1958, p. 497–501) have suggested a displacement of Africa with respect to Europe. The Triassic virtual geomagnetic poles from Europe have been cited in support of a rotation of Spain with respect to France and England (Clegg and others, 1957, p. 227). A 16-degree counterclockwise rotation of Japan has been suggested to explain Japanese results from Pleistocene to Holocene (Irving, 1959, p. 63). Finally, the large displacement of the Siletz River volcanic series' virtual geomagnetic pole from the usual North American polar-wandering curve has been suggested as possibly due to a large clockwise rotation of Oregon with respect to the rest of North America (Irving, 1959, p. 65).

Evaluation and interpretation of pre-Oligocene paleomagnetic data.—In analyzing the post-Eocene, and especially the post-early Pleistocene, paleomagnetic results, we found a dipolar field configuration and were able to make an estimate of the expected scatter in an individual measurement. With this information we could then discuss, with some confidence, the application of these paleomagnetic data to possible post-Eocene polar wandering and continental drift. In interpreting the pre-Oligocene data, is there any time for which we may also establish a field configuration and an estimate of the scatter?

The Permian and to some extent the Carboniferous are beginning to emerge as such times (Fig. 35) since there are now enough relatively consistent results from North America and Europe to establish an average pole region and to estimate the expected scatter. Even for these periods, however, the powers of resolution of the paleomagnetic method should not be overestimated. Displacements or rotations as small as 20°, such as have been suggested for Newfoundland (Nairn and others, 1959, p. 596), might well be due to expectable intraperiod variation rather than to displacements of land masses. A subsequent study of several Carboniferous formations from eastern Canada by Du Bois (1959b, p. B.A., 63) tends to confirm this conclusion.

Viewed on a larger scale, however, the consistency of the Permian and Carboniferous results from North America and Europe is most impressive and certainly indicates a magnetic-field configuration vastly different from the present configuration. If the earth's magnetic field was that of an axial dipole, as it probably was from Oligocene to early Pleistocene time and almost certainly was from late Pleistocene to Recent time, then these results constitute a strong case for polar wandering. This interpretation is also in accord with the axial-symmetry requirements of the dynamo theory.

An objection to this axial-dipole interpretation has been voiced by Öpik (1955, p. 236), however, who suggests that boundary conditions, such as temperature differences at the core–mantle boundary, rather than the earth's rotation might act to establish order in the convective core motions. The resulting external field would not be symmetrical with respect to the earth's axis of rotation, and, moreover, different rates of rotation of the core and mantle would not tend to give an average axial symmetry because the cold and hot spots causing the convective pattern would move with the mantle. This theory faces some obstacles, however. No analysis has, to our knowledge, been carried out to determine whether such temperature differences could exert a substantial influence on the convective pattern. Another obstacle to the theory is the axial-dipole nature of the earth's field for the past 30 million years. Either the temperature differences did not exist during this time, or they were fortuitously symmetrical with respect to the axis of rotation, or conceivably they were causally related to the rotation axis. Although the principle of uniformitarianism favors a polar-wandering interpretation for the Permian and Carboniferous

paleomagnetic results, Öpik's objection should be kept in mind as a possible alternative explanation.

probably not accurate to better than 20 million years, and this certainly appears to be a reasonable estimate since most of the formations

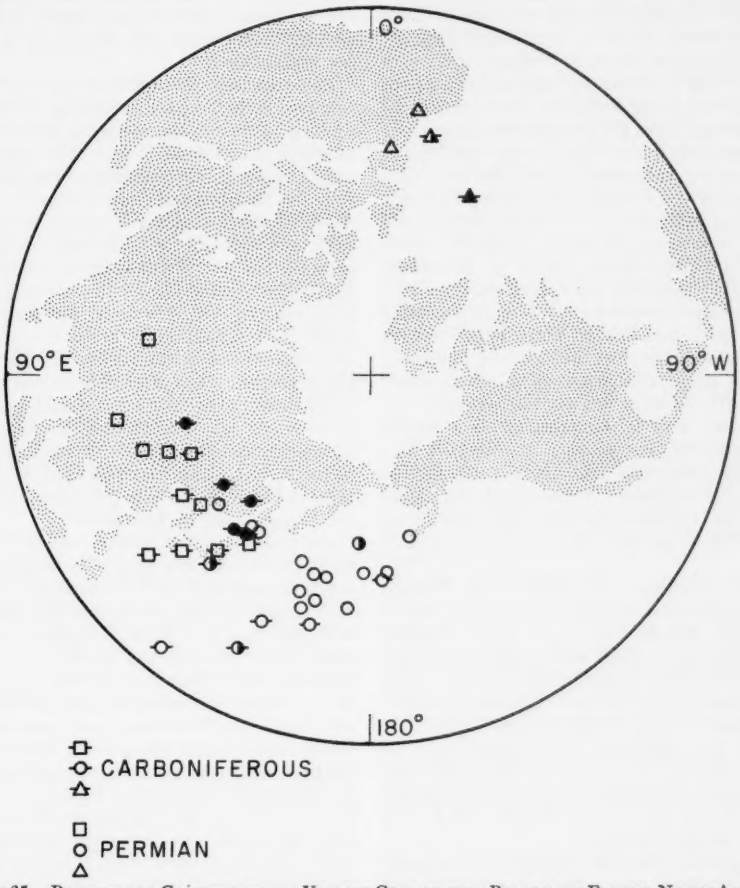

FIGURE 35.—PERMIAN AND CARBONIFERNOUS VIRTUAL GEOMAGNETIC POLES FROM EUROPE, NORTH AMERICA, AND AUSTRALIA
North America—squares, Europe—circles, Australia—triangles

The Permian data also constitute the strongest evidence for a relative displacement between North America and Europe. An alternative explanation for the displacement of the virtual geomagnetic poles calculated from the two regions may lie in inaccuracies in the stratigraphic correlation for the Permian on the two sides of the Atlantic Ocean. In discussing and rejecting this possibility, Creer and others (1957, p. 151) state that the correlations are

studied are lava flows, intrusive rocks, and continental red beds, all subject to more than average age uncertainty. Using an estimate for the average rate of movement of the dipole axis during the Permian (Creer and others, 1957, p. 155), we can compute that an apparent displacement of 10° would result from a 20 million year discrepancy in stratigraphic correlation.

Even assuming that there is no systematic difference in the age of the rocks from North

America and Europe listed as Permian, and assuming also that the field was dipolar, it is important to note that the two sets of Permian data from North America and Europe do not uniquely determine the relative positions of the two continents before a hypothetical displacement. The data from each continent must span enough time to delineate a segment along a path of polar wandering, and a very interesting problem emerges when the Permian and Carboniferous magnetic poles from Europe and North America are considered jointly. Between Carboniferous and Permian time the average geomagnetic pole position for North America appears to have moved roughly N. 85° W., while that for Europe moved roughly N. 40°–70° E. (Fig. 35). These two path segments cannot be brought into coincidence by postulating a simple westward movement of North America with respect to Europe since the Paleozoic, but would require, rather, a large post-Permian movement of North America *toward* Europe from the southeast. This is not intended as a serious interpretation, but points out difficulties that arise in attempting a unique solution for the data now available. (Geometrical techniques useful in experimenting with paleomagnetic reconstructions are given by Irving 1958, p. 227–229.)

The Australian Carboniferous and Permian paleomagnetic investigations, although few in number, include concordant results between sediments and volcanic rocks, with an excellent fold test for the sediments. However the virtual geomagnetic poles are quite significantly different from the poles for Europe and North America. If the axial-dipole hypothesis is valid for these periods, the Australian data constitute evidence for a relative displacement of Australia with respect to North America and Europe which cannot be ignored.

In general, there are relatively few measurements available from early Paleozoic rocks, and those that are available show a considerable range in pole positions (Fig. 23). Therefore, without sufficient data to determine a field configuration or to assess the expected scatter in the results, interpretations for these data involving continental drift or polar wandering are hazardous.

Many more Precambrian data are available, and, as might be expected from older rocks with large relative age differences, the scatter is large. As mentioned earlier, many of the virtual geomagnetic poles tend to lie in the eastern Pacific near the equator. Even if this group of poles represents an average field configuration, we must expect a rather large variation in any given measurement. Thus, as for the early Paleozoic, a large uncertainty exists in the basic data available for polar-wandering and continental-drift interpretations, and the paths shown in Figure 33 might best be regarded, at present, as working hypotheses.

Paleomagnetic measurements for the Mesozoic and early Tertiary present an interesting problem; although the preceding Permian and following post-Eocene results are each internally consistent, the geomagnetic pole positions from the Mesozoic and early Tertiary are quite scattered. Some impressively consistent results have been obtained for individual Triassic formations, and the distribution of results from North America and Europe to some extent suggests relative drift. However, the scatter between mean formation directions is quite large and weakens the conclusion that a relative displacement has taken place. The paleomagnetic results from the Jurassic are at least as scattered as those for the Triassic, and no conclusive statements concerning drift or wandering can be made. (Note also that poles **H5**, **H 7**, and **H 10** in Figure 30 are from formations of uncertain age and may be Jurassic.)

The paleomagnetic results from the Cretaceous and Eocene may profitably be considered together, inasmuch as large amounts of drift with respect to North America and Europe have been postulated for India, for Japan, and, to some extent, for Australia on the basis of geomagnetic data from rocks of these ages. Such interpretations are based on the conclusion that the Eocene and Cretaceous pole positions for North America and Europe were within 18° of the present geographic north pole and moved up to coincidence with it in the Tertiary, as shown in Figures 33 and 34. The paucity of paleomagnetic information for these time intervals has already been discussed. In this respect one of the most reliable virtual geomagnetic poles from the few available North American determinations (the Siletz River volcanic series) falls much closer to the Indian and Australian paleomagnetic poles than to the Cretaceous and Eocene points on the usual polar-wandering curves for North America and Europe. Thus, as for many of the other periods, the North American (and possibly European) field configurations for Cretaceous and Eocene time are far from certain, and, without such a well-defined reference position, the drift interpretations for Australian and Indian Cretaceous and Eocene results may be questioned.

An alternative explanation of the Cretaceous

and Eocene measurements, and one that does not require simultaneous rotations of several tens of degrees for Oregon and India, would involve relatively rapid changes in the magnetic-field configuration during this period while maintaining the present continental configuration. The field may have been nondipolar or, alternatively, may have been that of a dipole undergoing somewhat rapid changes in direction. Such changes may or may not have been accompanied by rapid polar wandering.

This hypothesis poses many problems. It appears to us, however, to fit the available data at least as well as the drift hypotheses and it emphasizes the uncertainties connected with paleomagnetic interpretations at this time.

We have emphasized the tentative nature of some paleomagnetic interpretations. However, we would not like to conclude without emphasizing the contributions of paleomagnetism to geology. The post-Eocene results are impressive and offer very strong evidence for the dynamo theory and against substantial Tertiary polar wandering or continental drift. Although the nature of the Mesozoic and early Tertiary magnetic field is obscure, the consistency of the late Paleozoic studies, however, indicates that the paleomagnetic method is quite applicable to older rocks. It seems probable, therefore, that, with additional carefully studied and well-dated determinations from Mesozoic and pre-late Paleozoic time, a clear picture of the field configurations will evolve, opening the way for better evaluation of the hypotheses of continental drift and polar wandering.

Directions of Paleomagnetic Research

In concluding this review, we wish to suggest some lines of investigation in geology and geophysics that are of special interest to the worker in rock magnetism, and also to mention some important problems that remain for the paleomagnetist.

Paleobiogeographic studies probably rank first among those in other disciplines that are of interest in paleomagnetism. The paleomagnetic data from post-Eocene rocks indicate that the average geomagnetic pole was close to the present geographic pole throughout this time, and additional independent paleobiogeographic evidence to show that the geographic pole has also been in its present position throughout this period would give added support to the axial-dipole magnetic-field theory. Moreover, large discrepancies appear in the paleomagnetic data of the late Mesozoic and early Tertiary, and relevant paleobiogeographic evidence to indicate whether these discrepancies could be due to continental drift or polar wandering is needed. The Permian and Carboniferous periods are paleomagnetically very calm, and the measurements indicate a vastly different field configuration from the present one. Thus, Permian and Carboniferous formations should also be interesting subjects for paleobiogeographic studies. Recent paleobiogeographic studies of some relevance are those of Durham (1950; 1952), Stehli (1957), and Irving (1956a).

A further development of the dynamo theory for the earth's magnetic field is also highly desirable. The question of greatest interest concerns the axial-dipole requirements of the theory; the paleomagnetist would like to know if it is at all possible for a nonaxial dipole field to exist for a long period of time. Could stresses, such as might be provided by a local "hot spot" in the mantle, cause an ordering in core motions that would result in a magnetic field not symmetrical with the axis of rotation?

Other applications of rock-magnetism techniques to special geologic problems might be mentioned. Since the thermo-remanent magnetization of some rocks acquired in each temperature interval is independent of that acquired in adjacent temperature intervals, an application to unraveling the thermal histories of these rocks may be possible. Magnetic-anisotropy properties of rocks may also find important geologic applications. Techniques that permit a rapid measurement of the preferred orientation of ferromagnetic grains having either flat or elongate shapes or magnetocrystalline anisotropy are now available.

A field of paleomagnetic research that has been merely alluded to in this paper is that described by Thellier and Thellier (1959) in a review of their extensive work on the intensity of the earth's magnetic field during historic and Quaternary time. Their report should certainly excite more interest in this subject. Although directions of the field are of most use in continental-drift and polar-wandering interpretations, intensity data should be very useful in developing an expanded theory for the origin of the field and, hence, of considerable indirect interest to studies of drift and wandering.

In order to test properly the hypotheses of large-scale continental drifts since the Cretaceous and Eocene, it will be necessary to have additional paleomagnetic data from well-dated Cretaceous and Eocene rocks from all continents, especially from Europe and North America. Rocks of Permian and Carboniferous

age from localities other than those already sampled in North America, Europe, and Australia should also be rewarding subjects for paleomagnetic studies.

In the paleomagnetic work completed, a relatively large number of studies from a given continent and interval of geologic time have been necessary before a clear picture of the field for that time and place has emerged; the Permian of North America and Europe are examples. Even with a large quantity of data, however, the picture may still be cloudy, as for example in the Triassic from North America and Europe. Future investigations may yield more and better information if they are able to satisfy the following conditions:

(1) The geologic formation to be studied should be well dated geologically and accessible at several sites over a relatively large sampling area.

(2) Careful consideration should be given to the sampling scheme so that a minimum of samples will be required, the requirements of the statistical method to be employed will be satisfied, and the earth's field over a large area and at several points in time is represented.

(3) Demagnetization and anisotropy studies of the samples should always be made in order to test the stability and paleomagnetic applicability of the measured magnetizations.

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RIVER MEANDERS

BY LUNA B. LEOPOLD AND M. GORDON WOLMAN

ABSTRACT

Most river curves have nearly the same value of the ratio of curvature radius to channel width, in the range of 2 to 3. Meanders formed by meltwater on the surface of glaciers, and by the main current of the Gulf Stream, have a relation of meander length to channel width similar to rivers. Because such meanders carry no sediment, the shapes of curves in rivers are evidently determined primarily by the dynamics of flow rather than by relation to debris load.

Velocity distributions along river curves provide a generalized picture of flow characteristics. Evidence on flow resistance in curved channels suggests that a basic aspect of meander mechanics may be related to the distribution of energy loss provided by a particular configuration or curvature. No general theory of meanders is as yet satisfactory, however; in fact, present evidence suggests that no single theory will explain the formation and characteristics of all meanders and that few of the physical principles involved have yet been clearly identified.

RÉSUMÉ

Dans la plupart des courbes de rivières la valeur du rapport entre le rayon de courbure et la largeur du lit, de l'ordre de grandeur de 2 à 3, est presque toujours la même. Les méandres formés par l'eau de fonte à la surface des glaciers et par le principal courant du Gulf Stream, montrent une relation entre la longueur du méandre et la largeur du chenal semblable á celle des rivières. Puisque de tels méandres ne transportent pas de sédiments, on peut en conclure que la forme des courbes de rivières est determinée principalement par la dynamique de l'écoulement d'eau plutôt que liée à la charge sédimentaire.

La distribution des vélocités le long des courbes de rivières donne une image générale des caracteristiques d'écoulement. Ce qu'on sait de la résistance à l'écoulement dans les chenaux courbes suggère que la résistance minimum est un facteur de base dans la mécanique des méandres. Toutefois, aucune théorie générale des méandres n'est encore complètement satisfaisante. En fait, les données actuelles suggèrent qu'aucune théorie ne suffira pas à expliquer tous les méandres, et que quelques-uns seulement des principes de physique en jeu ont déjà été reconnus.

ZUSAMMENFASSUNG

Die meisten Flußwindungen haben etwa dasselbe Wertverhältnis von Krümmungsradius zu Bettbreite, ungefähr zwei zu drei. Mäanderwindungen, die durch Schmelzwasser auf der Oberfläche der Gletscher entstehen, und solche, die durch die Hauptströmung des Golfstromes bedingt sind, haben ein Verhältnis von Mäanderlange zu Bettbreite, welches dem der Flüsse ähnlich ist. Da solche Mäander kein Sediment mit sich führen, sind die Formen der Flußwindungen offensichtlich in erster Linie durch die Dynamik der Strömung bedingt, weniger durch die Gegebenheiten des mitgeführten Materials.

Die Verteilung der Strömungsgeschwindigkeit entlang den Flußwindungen zeigt ein verallgemeinertes Bild der Strömungsmerkmale. Was über den Strömungswiderstand in gewundenen Flußbetten beobachtet wurde, läßt die Vermutung aufkommen, daß der geringste Widerstand eine wichtige Ursache für Mäanderbildungen ist. Tatsachlich lassen die gegenwärtig beobachteten Tatsachen vermuten, daß vorläufig noch keine einzige Theorie alle Mäanderbildungen erklaren kann, und daß nur wenige der physikalischen Gesetzmäßigkeiten, die dabei eine Rolle spielen, bisher gefunden worden sind.

МЕАНДРЫ РЕК

Люна Б. Леопольд и М. Гордон Уолмэн

Резюме

Большинство речных излучин обладает почти одинаковой величиной отношения радиуса кривизны к ширине канала, колеблющейся в порядке от 2 до 3. Меандры, образованные талой водой на поверхности ледников и

излучины главного течения Гольфштрема, обладают отношением длины излучины к ширине канала, подобным тому, которое наблюдается в реках. Ввиду того, что такие меандры не содержат осадков, формы излучин в реках, очевидно, определяются в первую очередь динамикой течения, нежели зависимостью от наличия кластического материала.

Распределение скорости течения вдоль речных излучин дает обобщенную характеристику течения. Доказательством сопротивления течения в криволинейных каналах наводит на мысль о том, что основным фактором в механике образования меандр является минимальное сопротивление. Еще не выработана общая удовлетворительная теория о происхождении меандр. В действительности, современные данные дают возможность сказать, что отдельная теория не даст объяснения происхождения всех типов меандр. Еще не изучены многие физические факторы, влияющие на происхождение меандр.

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General Statement

The most characteristic features of all stream channels, regardless of size, are the absence of long straight reaches and the presence of frequent sinuous reversals of curvature. These bends, whether or not they display regular reversals and sufficient symmetry to warrant the name meander, tend to be scaled versions of a given set of proportions. The proportions are determined by three dimensions: the repeating distance or length of the curve, the width of the channel, and the radius of curvature. In brief, the size of the bend appears to be proportional to the size of the river; large rivers have large bends, and small rivers have small bends.

An adequate theory of meanders should explain why all river bends seem to be of proportional sizes, and by what mechanical or physical principle a bend becomes adjusted in size to the stream discharge. Such a theory

should also explain why some rivers meander and others do not. It should explain why meandering rivers occur on ice with high velocity of flow and no sediment load, and why tortuous channels are so prevalent in tidal estuaries. A meander theory should also explain how chance perturbations of the flow may persist and lead to the development of a sinuous pattern.

The present paper does not provide any theory. An attempt is made to analyze and integrate what appear to be the most prevalent essential characteristics of meandering channels in nature. At least some of these prevalent characteristics are probably basic to the mechanics of meandering channels which are as yet poorly understood. These include the external features of the channel, as well as the distribution of velocity and the pattern of flowing water. Such a summary may prove useful in the eventual definition of the necessary and sufficient conditions for meandering of channels in diverse natural environments.

There is a large published record on the occurrence and physiographic setting of meandering channels, as well as on the form and theory of river curves. It is not our purpose here to cite this record exhaustively, but rather to sift out those facts which appear to us most pertinent, emphasizing at the same time that opinions may differ as to what is pertinent.

Our studies of meandering streams and of meander mechanics lead us to write a somewhat different kind of a paper than might usually be expected in a "review" series. Many new observations are available, some previously unpublished, some recently published or in process of publication. We hope it will be of some service, especially to the reader who has an interest but not a specialty in the subject, to have these new observations summarized and related to earlier observations. We believe that the status of the problem will be clearer as a result, and these observations are, in our opinion, of some importance in indicating directions toward which future work might profitably be directed.

We then attempt to show that the problem of river meanders can be broken down into two parts. The first concerns the mechanical and hydraulic processes that govern the form, size, and probably the occurrence of meanders. The second includes the physiographic history of a particular channel and the formation, maintenance, and possible dissolution of a meandering pattern. Although these two parts are almost inseparable they do represent quite different ways of viewing the meander problem. We hope it will be clear from this review that many important physiographic questions cannot be answered with satisfaction because there is insufficient knowledge at present about meander mechanics.

The authors are much indebted to J. Hoover Mackin and W. W. Rubey for many constructive suggestions and for a critical reading of an early draft of the manuscript.

Geometry and Patterns of Channel Bends

Because most channels are not straight but sinuous, the forms which they display in plan view are, in detail, infinitely varied. In our experience it is unusual for a reach of natural stream to be straight for a distance exceeding 10 channel widths (Leopold and Wolman, 1957, p. 53). Yet certain common features are discernible. These features, common to many individual bends, probably have some significance in the mechanics or hydraulics of meanders, but this significance is only poorly understood. The geometric characteristics of a channel seen in plan view are defined by Figure 1.

We use the term "arc distance" to mean the length measured along the channel center line from one point of inflection to the next. "Sinuosity" we use to mean the ratio of arc distance to half the meander length.

The geometry of meander curves has been the object of extensive statistical study, examples of which are papers by Jefferson (1902), Inglis (1937; 1949, Pt. 1, p. 143), Bates (1939), and Leopold and Wolman (1957, p. 58). An additional independent sample in the appendix includes measurements of rivers in which the presence of a well-defined flood plain indicates that the channel form and pattern have developed reasonably free of bedrock control. These data for rivers of moderate to large size were derived from published topographic maps and for small streams from plane-table maps made by the authors. The appendix also includes some data from published laboratory studies of meandering streams in erodible materials.

Studies of the geometry of the patterns of meandering rivers give generally similar results.

The most consistent correlations are between meander length, channel width, and radius of curvature, as shown by Figure 2, plotted from the data in the appendix. Because of the variability in the dimensions of channels in nature, considerable scatter of plotted data is expected. Despite the scatter the relations among the factors appear to hold through a very large range of stream size, from laboratory streams a foot wide to the Mississippi River nearly a mile wide. Various authors have expressed the relations for rivers in alluvial flood plains by regression equations, a few of which are presented in Table 1. Table 1 also includes relations derived from the data in the appendix.

The exponents in the regression equations are all so close to unity that the relations between meander length (wave length), amplitude, radius of curvature, and channel width may be considered linear. Among different groups of data fairly consistent values are obtained for the coefficient relating meander length to channel width; meander length ranges from 7 to 10 times the channel width. This ratio holds in general not only for meanders on a broad flood plain but also characterizes sinusoidal or meandering valleys incised into consolidated materials (Dury, 1958, p. 109).

A marked inconsistency is exhibited, however, in the ratio of amplitude to channel width. Amplitude correlates only poorly with meander length. We tentatively interpret these findings to mean the amplitude of

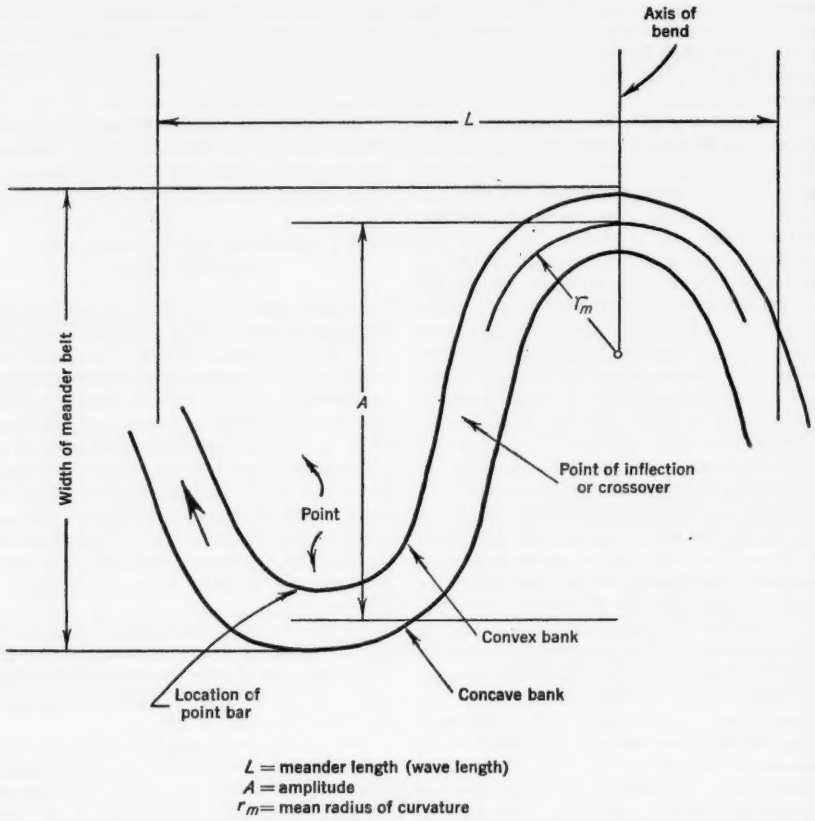

FIGURE 1.—SKETCH TO DEFINE TERMS USED IN DESCRIBING GEOMETRIC CHARACTERISTICS OF A MEANDERING CHANNEL

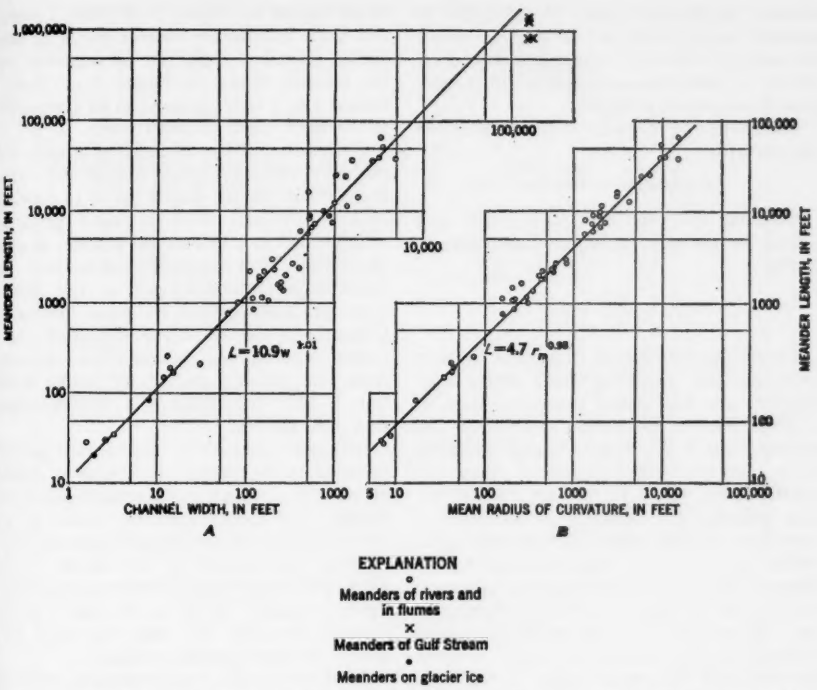

FIGURE 2.—RELATIONS BETWEEN MEANDER LENGTH AND CHANNEL WIDTH (LEFT DIAGRAM) AND MEAN RADIUS OF CURVATURE (RIGHT DIAGRAM)

TABLE 1.—EMPIRICAL RELATIONS BETWEEN SIZE PARAMETERS FOR MEANDERS IN ALLUVIAL VALLEYS

	Meander length to channel width	Amplitude to channel width	Meander length to radius of curvature
Inglis (1949, pt. 1, p. 144, Jefferson data)	$L = 6.6w^{0.99}$	$A = 18.6w^{0.99}$	..
Inglis (1949, pt. 1, p. 149, Bates data)	..	$A = 10.9w^{1.04}$	..
Leopold and Wolman (this paper)	$L = 10.9w^{1.01}$	$A = 2.7w^{1.1}$	$L = 4.7r_m^{0.98}$

meander loops is determined more by erosion characteristics of stream banks and by other local factors than by simple hydraulic principles.

The U. S. Waterways Experiment Station studies (Friedkin, 1945, p. 15) showed that in uniform material amplitude did not progressively increase nor did meander loops cut off as the meanders migrated downstream. Friedkin interpreted this to mean that "natural cut-offs result from local differences in the erodibility of bank materials (p. 16)."

These considerations appear to support the concept that a limit to the amplitude of meanders is provided by the formation of cut-offs. The rate of downstream migration relative to the rate of growth of the amplitude of bends probably depends on the distribution of shear against the stream banks, which in turn is a function of channel shape. The latter is determined at least partly by the characteristics of the flood-plain alluvium and the moving debris.

We believe relations between channel width,

meander length, and radius of curvature, in contrast to amplitude, are to a great extent independent of bed and bank materials and are related in some unknown manner to a more general mechanical principle.

Examining the correlations presented above one can write

$$L = 10.9w^{1.01} = 4.7r_m^{0.98}.$$

Assuming the exponents to be unity and solving for the ratio of radius of curvature to width,

$$r_m/w = \frac{10.9}{4.7} = 2.3.$$

Computing the value of r_m/w river by river from the data in the appendix rather than equating the least-square regression lines, it is found that in the sample of 50 rivers the median value is 2.7, mean 3.1, and two-thirds of the cases lie between 1.5 and 4.3. About one quarter of the values lie between 2.0 and 3.0. The tendency for similarity in the ratio of curvature to width makes all rivers look quite similar on planimetric maps, as suggested by Figure 3. In fact when one inspects a planimetric map of a river without first glancing at the map scale it is not immediately obvious whether the river is large or small owing to this tendency for similar ratio of radius to width regardless of river size.

As will be mentioned later, the value of the ratio of radius of curvature to channel width has an important bearing on resistance to flow, and this appears to offer some clue as to the hydraulic basis for the observed geometric similarity among channels of different sizes.

Curves in Sediment-Free Channels

Meandering channels are carved in glacier ice by streams of meltwater flowing on the glacier surface. To obtain measurements of such channels we went to one of the Dinwoody glaciers on the east side of the crest of the Wind River Range in Wyoming (Lat. 43°12′, Long. 109°38″). The meandering channels on this small valley glacier can be seen in Plate 1. Although concentrated primarily upon the steeper slope downstream, meandering channels began on the relatively flat portion of the profile, that is, near the point of inflection from concave to convex portions of the surface profile. The channels extended downstream to the convex-upward part of the glacier snout.

Some of the meandering channels were incised 6 to 12 feet into the ice, and those which appear on Figure 1 of Plate 1 are of this kind. Satisfactory measurements of these incised meanders could not be obtained, but the channel shown in Figure 2 of Plate 1 incised 1 to 2 feet, appeared to be comparable to the more deeply incised streams.

Comparison of the meandering stream with incipient channels being formed by water flowing over the ice surface led us to conclude that an irregular sheet of water gradually concentrated in a shallow flat channel at some place where local factors favored melting. The meandering form developed as the flowing water gradually melted a channel downward. Channels in the ice only a foot wide and 2 inches deep had no well-developed meander form, but another channel of similar width but 1 foot deep showed a well-developed sinusoidal pattern.

Although occasional rocks and pebbles occurred in the channel none were in motion at time of observation. In some bends a few pebbles had collected on the inside of the curve in the position of the point bar in ordinary streams. The surface of the glacier in the locality was strewn with cobbles and rocks in a random manner owing to the fact that the channel measured was near the edge of a poorly developed medial moraine.

In several places terrace remnants were cut into the ice and abandoned as downcutting progressed. No pebbles were seen on these terrace remnants. The meanders appeared to be cut by melting alone with no effective help from abrasion by transported pebbles.

Over-all dimensions (in feet) and flow parameters of the channel pictured in Figure 2 of Plate 1 were as follows:

Channel width	2.3
Channel incision below ice surface	1.0
Width of surface of flowing water	2.0
Mean depth of flowing water, d_m	0.25
Average velocity from measurements of flow, v	5.3 ft/sec.
Average meander length	20
Average channel slope, s	0.023
Discharge	2.4 cfs
Froude number $v/\sqrt{gd_m}$ (where g is the acceleration of gravity)	1.9

The Froude number, 1.9, was well within the range of supercritical or shooting flow. In shooting flow it is usual to observe surface wave trains and the diagonal crisscross of

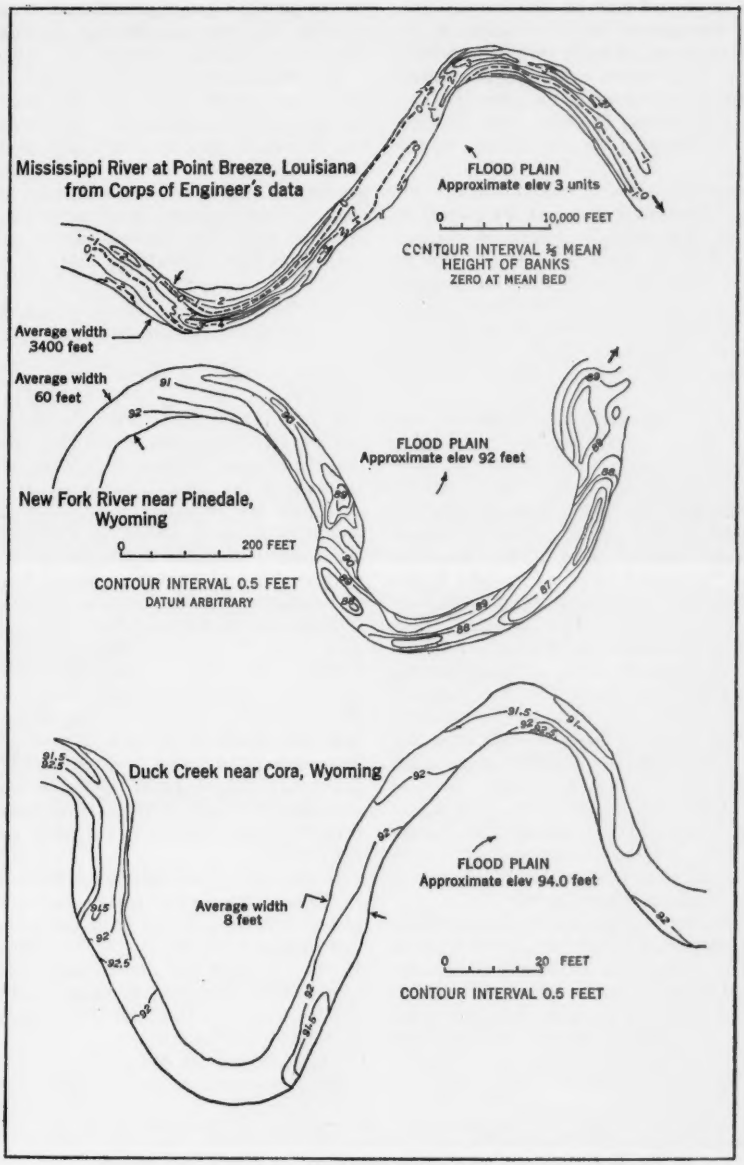

FIGURE 3.—PLANIMETRIC MAP AND BED TOPOGRAPHY OF A MEANDER OF THE MISSISSIPPI RIVER AT POINT BREEZE, LOUISIANA, THE NEW FORK NEAR PINEDALE, WYOMING, AND DUCK CREEK NEAR CORA, WYOMING

Scales are chosen so that meander length is equal on the printed page.

surface waves reflected off the channel sides. If these were present they were masked by the violent turbulence, although they were probably the cause of a marked hump in water-surface elevation extending along the channel center line in some places. The water was conspicuously superelevated on the outside of the curves.

The measurements of meander length and width of this stream cut in ice are plotted on Figure 2. The meander length of channels developed in ice bears the same relation to channel width as in ordinary meandering streams. Measurements made by D. G. Anderson of meanders of small streams on the ice of Chamberlain Glacier, Alaska, confirm this conclusion (Oral communication to Leopold, 1958).

Data on meanderlike phenomena from another source deserve mention as possibly having some relation to the present problem. In the Gulf stream of the North Atlantic oceanographers have found bands of relatively high speed which have a meandering pattern in plan view. Stommel (1954, p. 887) described these as "horizontal eddy and meander formations," and Von Arx (1952, p. 213) used the term "meandering current." These currents appear both in cross sections of the velocity field and in planimetric maps of the temperature field. Maps of the temperature field at a depth of 200 meters have been published by Stommel (1954, Fig. 2) and by Fuglister (1955, chart 3a). From some of these published maps rough measurements of meander length may be obtained and, less satisfactorily, estimates of current width. Better estimates of current width are obtainable from velocity cross sections published by Von Arx (1952, Fig. 2) and Worthington (1954, Fig. 9).

Four measurements of the width and meander length were possible from the published data, and these are plotted on Figure 2. Although they fall slightly below the line drawn through the points representing river data, the graph suggests that meandering currents in the Gulf Stream bear certain analogies to river meanders. Although frictional flow in rivers should not be compared to geostrophic flow of an ocean current, the observed similarity in meander dimensions appears worthy of attention.

The importance of meanders in ice and in ocean currents lies in the fact that both flow systems exhibit meanderlike phenomena in the absence of sediment debris. The establishment of meander length–width relations similar to those of sediment-laden rivers suggests that sediment alters or affects but does not cause the meander pattern. (For a contrasting view *see* Matthes, 1941.)

Channel Cross Sections and Longitudinal Profiles

From a study of bends of the Garonne River, Fargue (1908, p. 25) stated as a general rule of river behavior that the shallowest sections occur downstream from the crossover, and the deepest sections downstream from the axis of the bend or point of maximum curvature. The topographic maps of sample meanders in Figure 3 show in a general way the features described by Fargue and many workers after him. These same figures, however, also indicate that the location of deeps and shallows may be highly variable. For example, data from 25 bends on Buffalo Creek near Buffalo, New York, indicate that the deep points are usually found downstream from the point of maximum curvature but do occur as well upstream from this point (Parsons, 1959). On a reach of the Popo Agie River near Hudson, Wyoming, containing four successive bends of amazing symmetry deeps occurred upstream from the point of maximum curvature and not downstream.

In the reach of the Mississippi River shown in Figure 3 the deepest portion occurs at the axis of curvature in one bend but somewhat downstream from this axis in the other bend. In the New Fork the deepest portion in one bend occurs practically at the point of inflection. The reach on Duck Creek more nearly conforms to Fargue's generalization.

There seems little doubt that the depth of water at a bend of a river is closely correlated

Plate 1.—MEANDERS ON DINWOODY GLACIER, WYOMING

Figure 1.—In middle foreground are meandering channels of meltwater streams carved in the glacier ice. View southwesterly from Sentinel Peak; photograph by Mark F. Meier

Figure 2.—Channel carries meltwater from glacier. The water flowed at high velocity and was violently agitated.

FIGURE 1

FIGURE 2

MEANDERS ON DINWOODY GLACIER, WYOMING

with the bend curvature. Using 103 measurements of mean depth and corresponding radii of curvature on the River Elbe, Leliavsky (1955, p. 118) shows that the two factors bear a simple linear relation and that depth increases inversely as a function of radius of bend.

The point of inflection in a river curve is closely associated with a shallow portion of the reach, or a depositional bar on the bed. The occurrence of a bar or riffle in the bed of a straight river reach should correspond morphologically therefore to the point of inflection of a meandering channel. It is on this basis that twice the distance between successive riffles in a straight reach was compared with the wave length of meanders (Leopold and Wolman, 1957, p. 55).

Not only do the riffles in straight channels tend to be in a position analogous to comparable shallows in a meandering reach, but there also is some tendency for the shape of the bars to be the same. In some straight reaches successive gravel or sand bars which constitute the riffles assume the form of lobate wedges sloping alternately toward one bank and then the other. In some meanders the bar near the point of inflection slopes sharply across the channel toward the deep near the concave bank downstream. This is apparent on the topographic map of the New Fork (Fig. 3) and is accentuated at low flow when water flows diagonally across the channel down the steepest local bed slope. At high flow there is rapid divergence as the cross-sectional area expands.

The longitudinal profile of bars, whether associated with the point of inflection of a meander, or a riffle in a straight channel, is usually moundlike rather than dunelike and is asymmetric in cross section. Perhaps ideally the bar has a dunelike profile with a short, steep downstream face. The topography of the bend in the Mississippi River (Fig. 3) suggests this, but many are less regular.

The bar is asymmetric in cross section in a different sense on upstream and downstream faces. In a meander, as one moves downstream toward the point of inflection, the surface of the bar slopes laterally toward the concave bank, but immediately having passed the crest of the bar the asymmetry changes, and the surface of the bar slopes toward the opposite bank.

For this reason the channel cross section in a meandering reach is symmetrical only for a very short distance. We estimate that on the average the distance through which the cross section is essentially symmetrical is only about one-tenth wave length. This short distance can be seen on topographic maps of Figure 3.

The importance of the configuration of the bars and their relation to the deeps lies in the relation of bed topography to bed shear. River pilots and others know that in large rivers the crossings (points of inflection) tend to scour at low flow and fill at high flow; the bends (pools) tend to fill at low flow and scour at high (Straub, 1942, p. 617; Lane and Borland, 1954, p. 1075).

Local bed shear must be greater at points of scour than where filling occurs, but direct measurements of bed shear are few, and little is known in detail about the relation of bed topography to stress on the bed.

Lane and Borland (1954, p. 1079) explain the observed sequence of scour and fill as a consequence of the different shape of cross section in pool and riffle. They say that the crossing or riffle has a larger cross-sectional area than the pool during high stages and a smaller area during low stages. Laursen and Toch (1954, p. 1085) state that width often tends to be greater at a crossing than at a bend. The expansion in width they believe would cause deposition on the wide crossing while scour occurs in the narrower bend.

We have made measurements of slope and of depth over pool and riffle separately at various stages of flow in an attempt to compute approximately the variation of mean stress. At low flow the bend is a still pool over which the water has a relatively flat slope. In contrast, in the riffle the water tends to plunge over the steep downstream face of the bar. As stage increases, the slope over the pool or bend increases, while that over the bar decreases until near bankfull stage when the water surface attains a continuous uniform slope, and all obvious surface effects of the shallow bar have been erased.

With increasing stage, depth increases in both bend and crossing. The product of depth and slope, then, increases with stage more rapidly in pool than at the riffle or crossing.

In the one stream measured by us, Seneca Creek near Dawsonville, Maryland, the computed mean shear is greater in the riffle at low flow and greater in the pool at high flow (Leopold, in press). This observation is in qualitative agreement, then, with the concept of scouring in the crossing at low flow and in the bend at high flow. The analysis is incomplete, however, inasmuch as variations in the distribution of velocity in successive cross sections have not been included in the computation of shear, and additional observations

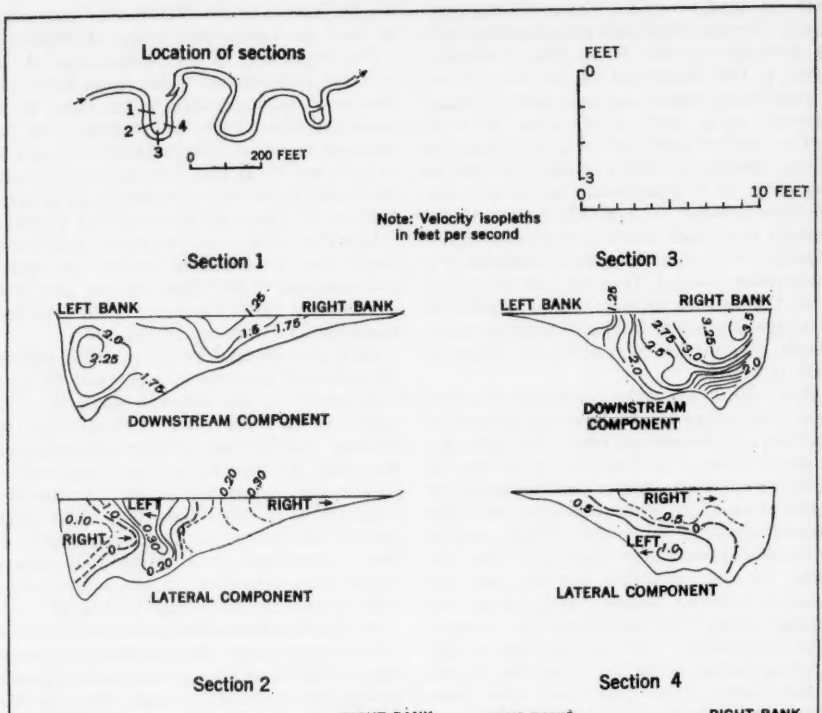

FIGURE 4.—LATERAL AND DOWNSTREAM COMPONENTS OF VELOCITY AT VARIOUS CROSS SECTIONS IN A BEND, BALDWIN CREEK, NEAR LANDER, WYOMING

elsewhere are necessary before any generalization can be made. Mean depth and slope, even over a short reach, probably do not measure the true shear exerted on any local area of bed. Direct measurement of shear stress in various parts of a bend at various stages is needed.

DYNAMIC AND FLOW CHARACTERISTICS

Pattern of Flow in a Meander Bend

Relatively few detailed measurements of the distribution of flow in a natural meander have been published (Blue *et al.*, 1934; Eakin, 1935; Leliavsky, 1955, p. 96–100; Van Til and Tops,

1953, p. 19–23). The velocity distributions which follow are intended both as additions to the slim store of data and as illustrations of the general characteristics of flow in bends in natural channels. Figure 4 shows downstream and transverse components of velocity at the axis of the bend on Baldwin Creek near Lander, Wyoming. Point velocities were measured by Price current meter and horizontal angles by a vane attached to the wading rod. Measured angles were used in conjunction with the measured velocity to compute the lateral components of velocity. Measurements were made at a high but less than bankfull stage.

At least in narrow channels a cross-channel velocity component is directed toward the convex bank or point bar near the bed, and toward the concave bank near the surface. Continuity requires, then, that surface water plunge downward near the concave bank and that some bed water emerge at the surface near the convex bank.

This circulatory motion in the cross-sectional plane of a channel, as clearly explained by Thomsen (1879), results from the larger centrifugal force on fast-moving surface parcels than on slower-moving ones near the bed. The motion gives to an individual water parcel a path resembling a helix.

Despite the tendency for circulatory motion in the stream cross section, surface stream lines (Fig. 5) show that no single parcel of water crosses completely from the convex to the concave bank. The maximum cross-channel motion does not exceed perhaps two-thirds of the channel width in any given meander bend. This accounts for the observation by Friedkin (1945, p. 5) in laboratory models that material eroded from one bank tends to deposit on a point bar downstream on the same side of the stream.

Matthes (1941, p. 634) stated that in wide, shallow rivers helical circulation does not exist. It is our opinion that in rivers of large width/depth ratio the helical motion exists, but not as a single rotating cell involving the whole channel (Nemenyi, 1946). The existence of several cells of circulation in the Tamm Bend data from the Mississippi can be seen in the distribution of the lateral components of velocity published by Eakin (1935). As in the atmosphere, circulation cells of large magnitude tend to break down into smaller cells. The "polar-front" cell and the "trade-wind" cell of Rossby (1941, p. 610) are directly driven by dynamic forces which we consider analogous to the direct helical motion in river bends due to the combined effect of frictional and centrifugal forces. Rossby's "frictionally driven" circulation cell in the middle latitudes furnishes, in our opinion, an analogy to those circulation cells seen in the Eakin data of the Mississippi which rotate in a direction opposite to the helical motion expected in a single cell occupying the entire cross section.

It is well documented that the water surface is superelevated near the concave bank of a channel bend (Blue *et al.*, 1934; Eakin, 1935; Leliavsky, 1955, p. 123; and others). The amount of the superelevation is proportional to

$$\frac{v^2 w}{g r_m}$$

where v is the mean velocity, w is channel width, r_m the mean radius of curvature, and g the acceleration of gravity (Leliavsky, 1955, p. 122).

This superelevation is a consequence of the curved path of water flowing around a bend and is a correlative of the helical motion.

Generalized Picture of Flow in a Meander

Measurements in meandering streams and in curved flumes (for example Mockmore, 1944, p. 569) allow the construction of a generalized picture of the flow pattern in a meander (Fig. 5). The isometric view of the two principal components of velocity at various positions in the bend show the main features. The scale is such that superelevation of the water surface in the bend does not show on the diagrams but is implied by the velocity distribution.

At the crossover or point of inflection of the bend (Fig. 5, section 1), the cross-sectional shape is not completely symmetrical but is slightly deeper near the bank which was concave in the bend immediately upstream. Downstream from the crossover the section (2) becomes approximately symmetrical, but, because this section is in the bend, some cross-current component exists in accord with the curving stream lines.

The velocity in a meander crossover is not symmetrically distributed. As would be expected, proceeding downstream from the axis of the bend the thread of maximum velocity is much closer to the concave bank than to the center of the channel. The high velocity, moreover, continues to hug this side through the point of inflection of the curve.

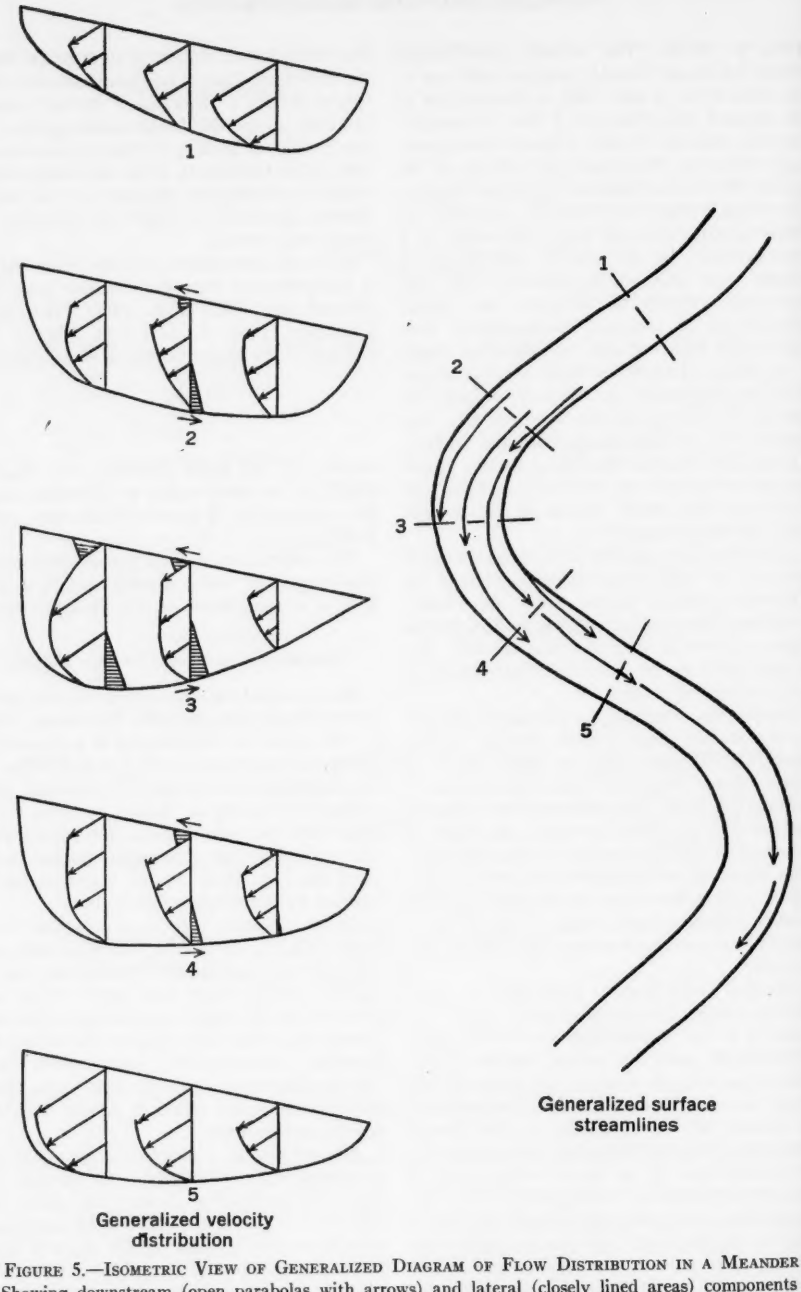

FIGURE 5.—ISOMETRIC VIEW OF GENERALIZED DIAGRAM OF FLOW DISTRIBUTION IN A MEANDER

Showing downstream (open parabolas with arrows) and lateral (closely lined areas) components of velocity as vectors, and surface stream lines. All sections viewed from a changing position to the left of and above the individual section.

The asymmetry in the velocity field persists downstream from the point of reversal of curvature, at least as far as the most nearly rectangular and symmetrical channel cross section.

Downstream from the symmetrical cross section the maximum downstream velocity approaches the concave bank. The highest velocity at any point occurs near the concave bank just downstream from the axis of bend. Individual filaments of water accelerate and decelerate along a stream line. The maximum velocity is not always associated with the same parcels of water.

Measured distributions of velocity show also that the maximum-point velocity in the bend is depressed below the water surface. These observations agree with the measurements made by Mockmore (1944, p. 596) in a curved flume.

An effect of increasing discharge in a meander reach is to move downstream the position on the concave bank where the thread of maximum velocity impinges. The stream lines of highest velocity do not touch the concave bank but approach it and then swerve into a course more or less parallel to that bank.

Channel Shape and Movement in Relation to the Flow Pattern

The slight lack of congruence of stream-line curvature with bank curvature leads to the tendency for the locus of point-bar deposition to occur downstream from the axis of the bend. A building point bar tends to concentrate caving of the bank downstream from the axis of bend and thus accounts for the progressive down-valley movement of meander bends. This same tendency exists in all river curves, even in those lacking sufficient symmetry to be called meanders, and this accounts in part for normal movement of river channels and the consequent construction of flood plains.

Material slumping into the bed by bank caving is caught in the transverse component and carried toward the middle of the stream. If the location from which it is derived is far enough downstream in the bend, insufficient length of reach remains for the material to get all the way across the channel. Instead it moves through the crossover reach without having crossed the channel. Once into the reversed curve the particle moving near the bed is drawn toward the same bank from which it started rather than to the opposite bank. This does not rule out the likelihood that much fine material which is not carried near the bed of the stream will be deposited at a downstream location on the side opposite to the one from which it was derived.

The plunging or diving motion of surface water near the concave bank is one factor tending to depress the point of maximum velocity below the surface near the axis of the bend. Bank friction also has a similar effect. Because the highest velocity gradient on both bed and bank occurs where the thread of highest velocity is nearest the boundary, this maximum velocity gradient is equivalent to maximum shear stress. Its location accounts, then, for deepening of the bend reach near the concave bank. This scour, together with building of the point bar, explains the triangular shape of the cross section.

Increasing the depth of water in the channel reduces the vertical velocity gradient. In a channel that is very deep relative to its width, the effect of bed friction becomes relatively small, and the velocity is nearly constant with depth except close to the bed. Thus the helical circulation becomes negligible. Flow in a curved channel of great depth and small width approaches potential flow—that is, the downstream velocity varies across the channel inversely as the radius of curvature of the stream lines. This was demonstrated by Wattendorf (1935, p. 574) in a deep, narrow, curved flume. In the absence of the circulation in cross section the thread of high velocity hugs the convex bank and does not cross the channel. Thus the mechanism for building the point bar is probably absent in channels of extremely small width–depth ratio.

Effect of Flow Pattern on Deposition and Erosion

The vigorous cross currents near the bed can transport considerable quantities of bed material from the concave to the convex side of the channel. On a meander bend of Baldwin Creek (Fig. 4) depth-integrated samples of suspended load were taken at each of five verticals spaced across the stream and repeated in each of five cross sections along a meander wave. A nearly uniform concentration of sediment existed across the section in a crossover. In the sections located in the curved reach the concentration increased markedly near the convex bank.[1] A similar distribution of sediment was observed by Eakin (1935,

[1] Dr. John P. Miller assisted the senior author in obtaining these measurements.

p. 471) in his samples taken at Tamm Bend on the Mississippi River. This increased concentration near the convex bank is attributed to the fact that water in this location has, to some extent at least, come laterally in a path close to the bed and is comprised, therefore, of water elements with relatively high concentrations of sediment.

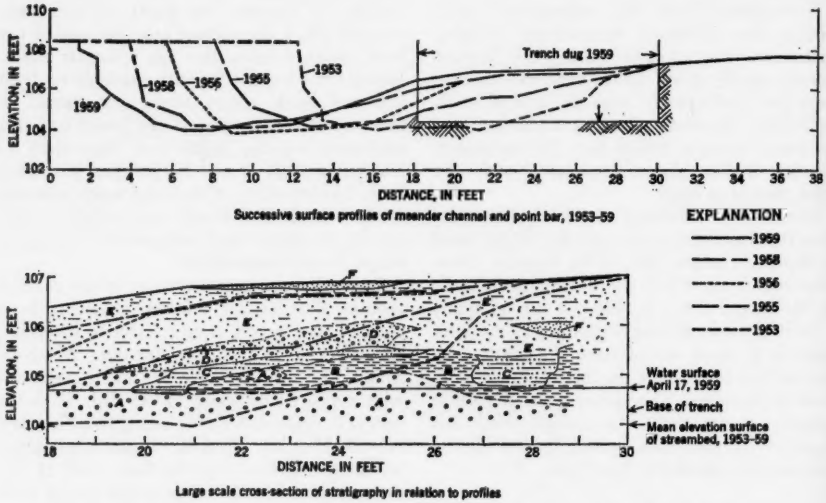

FIGURE 6.—STRATIGRAPHY OF POINT BAR IN RELATION TO MEASURED CROSS-SECTIONAL PROFILES DURING SIX YEARS OF OBSERVATION, WATTS BRANCH, 1 MILE NORTHWEST OF ROCKVILLE, MARYLAND

Key to stratigraphic units is as follows:
A Gravel, mostly 3–8 mm with considerable 8–20 mm and a few 64 mm
B Olive-gray clayey silt, with organic matter, small mica flakes
C Orange-brown, mottled sandy silt with some clay, lenses of leaflike organic mterial, mica
D Coarse sand, brown-stained, with pebbles up to 8 mm, some fine roots *in situ*, lenses of silt
E Brown sandy silt gradually changing upward in places into fine sand
F Fine sand with some silt

Although there is some evidence that rifling in a spirally welded pipe tends to increase the ability of a flow to transport sediment over that of a cast-iron pipe (M. L. Albertson, oral communication to authors), evidence from preliminary experiments (L. M. Brush, oral communication to Wolman) indicates that transport in a meandering channel is much reduced over that in a straight channel of similar dimensions and boundary roughness. This effect of curvature on transport is in accord with the added energy losses accompanying increased curvature.

The cross-channel component of bed flow toward the convex bank is probably the principal mechanism for the building of a point bar. The path of particles which are deposited on a building point bar has, of course, only a small component across the stream compared with a large down-stream component. The lobate form of a point bar may be compared to a sector of a very flat cone. A bed particle, taking a path along a chord of this cone, may climb the inclined plane or ramp presented by the cone. Thus, even some of the coarser debris of the channel bed will be deposited on the point bar at an elevation considerably higher than the average level of the stream bed.

Variation in discharge would lead one to expect that a thin layer of coarse material may be deposited on the point bar at one time and at another time a layer of fine material may be deposited upon the coarse. On the average, fine material will be carried higher up the bar than will coarse material, so the point bar would show on the average a rough gradation of size from coarse near the base to fine near the flood-plain level. In many instances, then, a growing point bar would not merely lap over the coarse material on the stream bed.

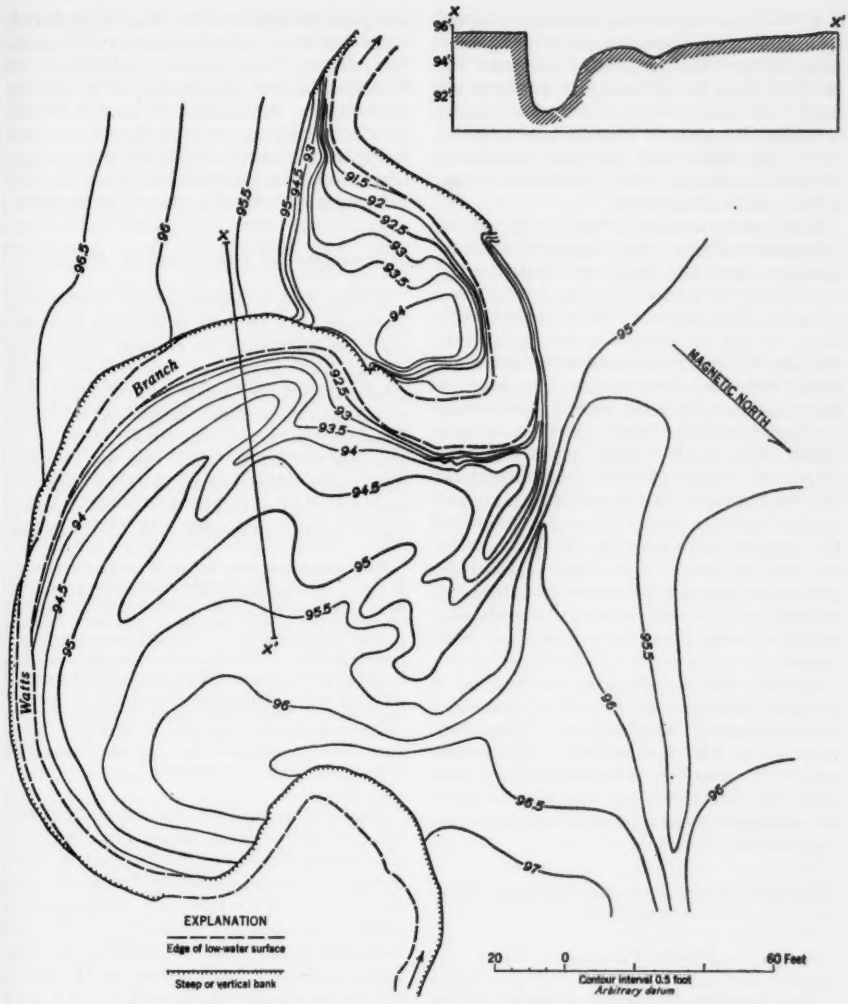

FIGURE 7.—MAP AND CROSS SECTION OF A TYPICAL POINT BAR OF THE FLOOD PLAIN OF WATTS BRANCH, 1 MILE NORTHWEST OF ROCKVILLE, MARYLAND
After Wolman and Leopold (1957)

An unusual opportunity to verify this reasoning was provided by a series of cross sections made in a meandering reach of Watts Branch near Rockville, Maryland. During 6 years of observation the stream has moved laterally a distance equal to one channel width. At the end of this period of observation a trench was dug in the point bar normal to the channel. The stratigraphy revealed in this trench superimposed on known surface positions is sketched in Figure 6. This trench was located on the right side of the stream on section x-x′ of Figure 7.

Approximate contact surfaces between materials of different textures are more or less parallel to past surface profiles. Coarse material characteristic of the bed is observed in a position well above the channel bed. In addi-

tion, there is no doubt that deposition occurred up to the level of the water surface at bankfull stage. In the same period deposition over the flat flood plain by overbank flow has been too small to measure. These observations appear to confirm the authors' (Wolman and Leopoldt 1957) hypothesis that point-bar building i, the primary process of flood-plain developmens in flood plains of this type.

Bank erosion is greatly influenced by wetting of the bank materials. Arroyos cut in fine-grained alluvium experience most bank cutting after, not during, flow. The wetting causes later slumping (Leopold and Miller, 1956, p. 5). Bank erosion is also enhanced by return seepage of water which infiltrated the banks during high flow. Upon lowering the stage the balancing pressure of the water in the channel is released, and the banks slump or collapse (Inglis, 1949, pt. 1, p. 152). A study of bank cutting in Watts Branch near Rockville, Maryland, showed that a combination of bank wetting and ice-crystal formation promoted the greatest bank erosion (Wolman, 1959, p. 214). Although the largest discharges occurred in summer, the winter provided more thorough soil wetting which, in combination with freeze and thaw, led to maximum bank erosion.

As bank erosion occurs in the bend of a meander, over a period of time it is usual for an approximately equal amount of deposition to occur on the opposite bank. This general equality of deposition and erosion is the reason width and cross-sectional area remain about the same as the channel moves laterally across the flood plain.

Meander Mechanics and Physiographic Problems

General Discussion

The vagaries of nature provide endless opportunities for perturbations in the flow—local bank erosion, chance emplacement of a boulder, fallen trees, or blocks of other vegetation—any one of which would alter the path of a straight channel. Thus one need hardly inquire why a stream channel is not straight. On the other hand, a random succession of chance perturbations might be expected merely to result in random bends of different patterns. Although this situation describes many channels, the existence of beautifully symmetrical meander bends and the remarkable similarity of bends in rivers of different sizes and physiographic settings must be explained. There is not yet available any theory or dynamical principle which explains qualitatively the characteristic patterns common to meandering channels. In the absence of such a general principle, however, attempts have been made to explain at least qualitatively how symmetrical successive bends begin and grow, and how size of bends is related to the size of the river.

Initiation and Development of Meanders

A large body of experience and literature on river regulation has been built up by European engineers on rivers in Europe, India, and Africa. This experience has covered a far greater span of time and of field conditions than has American practice. Regarding meanders, Leliavsky recently summarized the salient concepts developed (1955, esp. p. 111–141). Apparently the consensus of these workers is that the effect of helical flow is the dominant factor. Leliavsky expresses it (p. 128) as follows:

> "For some reason or other, a small abrasion ... in the ... bank of a straight channel is supposed to have taken place. The water particles moving alongside the eroded portion of the bank follow a curved trajectory and develop, consequently, a centrifugal force. This force, in turn, gives rise to a local helicoidal current, which intensifies the original abrasion and works its way deeper and farther into the shore, ... until the whole channel becomes finally involved in the process and the eroded material accumulates on the opposite bank. This, then, is the birth of a meander."

Prus-Chacinski (1954) also argues that helical flow is the basic mechanism leading to meandering. He showed, further, that, by introducing an artificial secondary flow at the entry to the first bend, it is possible to produce various kinds of secondary circulations in the next successive bend which, in turn, affect the circulation in the next bend, and so on. He demonstrates that the downstream effects of a given circulation pattern are quite persistent, often through several successive bends. The "cause" of meandering Prus-Chacinski ascribes to any disturbance which produces an initial secondary circulation.

It seems clear that helical flow may play an important role in the process of deposition on a point bar. A building point bar helps concentrate shear against the concave bank and thus promotes bank caving and channel movement. Even denying that helical flow exists in wide rivers, Matthes' concept of meander development is closely allied with that just mentioned.

Matthes (1941, p. 633) indicates that bank cutting and orderly transfer of sediment to its place of deposition on point bars is a principal requirement for meandering. He observed that material tends to be deposited on point bars on the same side of the channel as that of the concave bank from which it was eroded. Thus, by bank erosion of a concave bank and concurrent deposition on a point bar across the channel, the channel will move laterally and downstream. Thus, the meander wave may migrate downstream generally maintaining its configuration if the materials are uniform. Friedkin's concept of the process (1945, p. 4) is essentially similar to that of Matthes (1941).

Only a few observers have studied this process of meander formation and development in laboratory channels. Quraishy (1944) described the development of a sinuous channel from an initially straight one in a sand-bed flume. Of particular interest was the formation of a series of small dunelike ripples of sand which he called "skew shoals." These developed on the initially flat bed after the sand grains had been in motion for a short time. These skew shoals were regularly spaced and alternated in position along the channel sides. After their full development they obstructed the path of water and forced it to assume a sinuous course.

This kind of mechanism is highly suggestive because it produced alternating bends, spaced uniformly along an initially straight channel at distances depending on the spacing of depositional barlike features on the bed. Although the Quraishy experiments do not indicate how the shear force of flowing water interacts quantitatively with bed debris to govern size and spacing of the dunelike features, it implies that the geometry of meander waves might be a function of bed forms.

The coincidence of the spacing of pools and riffles in straight reaches and meander lengths in rivers of comparable size might well be the result of a principle of bar or dune formation, as yet unknown, allied in some manner to skew shoal formation. Wolman and Brush (in press) found, however, that in noncohesive sands similar skew shoals formed only at supercritical flow (Froude number, based on mean depth and velocity, greater than 1.0). Although their wide shallow channels did not meander, Friedkin (1945, p. 4) produced a meandering channel from an initially straight one and concluded that bank erosion alone was sufficient to initiate meandering as long as channel widening and shallowing did not proceed too rapidly. In the study by Wolman and Brush (in press) helical flow was observed in the channel bends produced from the skew shoals.

Werner (1951) among others[2] has attempted to develop a general equation to describe meander formation and geometry. He (1951, p. 899) has suggested that the initiation of meanders is caused by "some local impulse or disturbance" which creates a transverse oscillation in the straight stream channel. Expressing velocity in terms of roughness, slope, and depth, as given by the Manning equation (p. 900), he derives an equation in which meander length is a function of initial width, channel roughness, slope, depth, and a coefficient proportional to sediment load.

An assumption is made that sediment prolongs the period of oscillation, inasmuch as any mechanism relying principally on inertial oscillations of the water within the channel width (seiche effects) gives wave periods much too short to account for meanders. The equation agrees qualitatively with respect to the interrelationships of some parameters. It indicates, for example, that meandering will be inhibited at high slopes, a conclusion which appears to be borne out in natural rivres (Leopold and Wolman, 1957, Fig. 46). On the other hand, the equation does not appear to describe quantitatively the observed relation of meander length to channel width. Further, if sediment load is not a necessary condition of meandering, the equation is invalid, for, when the load is zero, meander length is zero, and hence meandering should not occur.

Hjülstrom (1942, p. 252; 1949, p. 84) presented an expression for meander length based on seiche theory. He computed wave length as a product of seiche period and wave celerity, period being that of a seiche having a fetch equal to the width of the meander belt and water depth equal to that of the river. Wave celerity he considered to be the mean downvalley velocity of the river. The wave length so determined was also made a function of a coefficient of turbulent friction.

The resulting expression provided Hjülstrom a basis for reasoning about the effects of changing discharge on wave length through the interacting effects of water depth, velocity,

[2] The Coriolis force or the lateral deflection due to the rotation of the earth has frequently been cited as the cause of meandering, but calculations by Ludin (1926) and others indicate that the virtual radius of curvature attributable to the Coriolis force in a stream with a velocity of 3 feet per second would be on the order of 8 miles at a latitude of 60°.

and turbulence, and his argument showed keen insight and understanding of field conditions (1949, p. 86–88). But the formula has the disadvantage of making meander length dependent on wave amplitude; in our opinion, measurement data do not demonstrate this.

Although a simple comprehensive expression is still wanting, it appears that the forces determined by the velocity distribution, including the helical circulation, are all that is necessary to account for (1) the shape of the cross section in a meander, (2) the depositional and erosional pattern, and (3) the progressive down-valley migration of the meander. These observations are not new, but it is important to emphasize the following idea. Although point-bar formation and associated erosion of the opposite bank are necessary if a straight channel is to develop curves, the concept of helical flow, as Leliavsky (1955, p. 128) recognized, does not seem to explain how the secondary circulation determines the characteristic dimensions or proportions of meandering channels. The existence of meanders on glacier ice also implies that erosion and deposition may be a collateral, not the governing principle of meander development and movement. Because the hydraulic or mechanical significance of the pattern of curvature is closely tied to the fundamental physiographic questions, we consider this aspect of meander mechanics in the following section.

Problem of Channel Equilibrium

The preceding discussion was concerned principally with the initiation and development of a meander, elements that can be discussed in terms of a short reach of river or a single meander wave. There are a host of broader problems of channel adjustment to external controls which might be thought of as physiographic problems for want of a more specific term.

To begin, one might ask how meandering of a channel relates to the fundamental process of stream adjustment and stream equilibrium. It is generally believed that channel equilibrium is constantly approached, although rarely attained, by a process of continual adjustment. To use the words of Rubey (1952, p. 129),

> "....with changing conditions, the stream is constantly cutting or filling and modifying its slope, velocity and cross section so as eventually to accomplish the imposed work with the least expenditure of energy."

If, indeed, a principle of least work is involved—for it is not yet proven—how does the development of a meandering pattern help accomplish this objective? Is the reduction of gradient achieved by increasing length relative to a straight channel necessary because of an excess of energy? Or, as was once believed, is meandering the aimless wandering of a channel too sluggish to accomplish any work of erosion? Does a river reach a stage at which vertical erosion is negligible and thence expend its excess energy on lateral erosion by meandering?

Under what changes of conditions would a river change its pattern from meandering to nonmeandering or vice versa? What would be the effect of an increase or decrease in discharge or in sediment load from the drainage basin?

Opinions on some of these questions have been published, but data or measurements are meager. Existing data may answer a particular question but do not explain why the observed result was obtained. In the following paragraphs some of these questions are considered along with related observations from the literature. Results of recent work, where applicable, are cited, and we suggest what appears to us some of the directions in which future work is needed.

Many of these queries relate to a single, fundamental question—what has the meandering pattern to do with energy expenditure? Inglis (1949, p. 158) states that,

> "Meandering is Nature's way of damping out excess energy during a wide range of varying flow conditions, the pattern depending on the grade of material, the relation between discharge and charge (load) and the rate of change of discharge and charge."

Schoklitsch (1937, p. 149) earlier stated what appears to be the same idea, that meander formation

> "might be due to the fact that the slope in such stretches is too great and is not in equilibrium with the size of the bed-sediment grains."

Water-surface slope of a river is a measure of energy expenditure. Whereas the equilibrium slope is closely related to the size of particles on the bed, there are other factors involved. Hack (1957, p. 61) confirms quantitatively the generally held belief that channel slope is controlled to a great extent by bed grain size and discharge, but to what extent channel shape enters is still unclear. As Rubey explains

(1952, p. 131), channel shape may adjust jointly with slope. At constant discharge if the principal effect of bed particle size is on channel roughness or frictional drag, then other forms of drag need to be considered. Total channel resistance is materially influenced by form drag of various kinds, pools and riffles, bars and dunes, and channel curvature.

There is reason from the hydraulic standpoint to believe that meandering may in part be a function of frictional drag and thence energy loss. More energy loss occurs per unit of length in a curved than in a straight channel of the same depth and cross section, owing to eddying, secondary circulation, or increased rate of shear. These eddy losses result from deflection of the water to a new direction as it moves around a curve or bulge. It is known from hydraulic experiments in pipes (King, 1954, Fig. 87) that energy loss first decreases and then increases with a decrease in the ratio radius of curvature/pipe diameter. Although any curved channel offers greater resistance to flow than a straight one, the minimum increase in resistance is about 40 per cent, and this applies within a narrow range of the ratio of radius of curvature/diameter when that ratio has a value of 2 to 3.

It was pointed out earlier that in meandering channels the comparable ratio, radius of curvature/channel width, is relatively conservative. Values of this ratio also tend to fall in the range of 2 to 3.

Bagnold (in press) suggests an explanation of this minimum increase in resistance at this value. He postulates that, as radius of curvature is reduced and becomes about 2 to 3 times the channel width, an eddy or zone of reverse flow forms just downstream from the bulge or convex bank. With the appearance of the reverse eddy the local width is constricted, and there is a local increase in effective radius of curvature and a net decrease in energy loss.

To the extent that further work confirms the indication that the modal value of this ratio is in the range 2–3, meanders tend to be characterized by a geometric pattern which happens to offer the smallest energy loss of any configuration of curved channel. The significance of this observation is unknown, but it suggests that some principle related to energy conservation does operate in the meander mechanism.

Assuming that a bend actually does tend to develop a configuration such that the energy loss due to the bend is a minimum, it does not follow that the total energy expenditure has been minimized. When a given discharge falls through some specified vertical distance, a certain amount of energy is transferred from potential form to some other form. If the water does not accelerate (velocity remains about constant downstream), then this potential energy is expended as work or heat. The same amount of energy is spent whether the water moves in a straight channel on a steep slope or in a longer curved channel at a smaller gradient. The energy expenditure per foot of channel length is smaller in the longer curved channel than in the shorter straight one. The question, then, is how much energy is utilized per foot of channel length and in what form it is used.

This energy may be spent in moving particles of debris, or it is otherwise dissipated into heat. The energy may be spent in removing particles from the bank and transporting them (bank erosion) or in transporting bed or suspended particles. If this energy is concentrated in such a manner that more of it goes into moving particles from one place than from another, then local scour will occur there and deposition elsewhere. For a channel to be in equilibrium scour must balance fill within the reach in question, and, further, the energy must be so expended that the net amount of debris coming into the reach must equal the net amount carried out of the reach.

It is generally agreed that meandering channels are often stable or in quasi-equilibrium. They may be so even though, over a period of time, a meander wave moves gradually downstream. The slope, discharge, and channel shape tend to become adjusted so that the above requirements are fulfilled. Adjustments in channel shape occur through erosion or deposition which in turn affect velocity, depth, and width. Specific hydraulic requirements relating depth, slope, velocity, and total resistance, including resistance offered by bed configuration (form resistance), bed and bank grains (skin resistance), and channel curvature (a particular kind of form resistance) must also be maintained.

Meandering is one way in which erosion and deposition may change the distribution, location, and amount of energy expenditure per unit of channel boundary. By lengthening the channel between two points at different elevations, the energy expenditure per foot of length is reduced. By bank erosion, point-bar building, and by scour and fill, the channel

cross section is adjusted, and the energy expenditure is redistributed. Presumably abrupt discontinuities in the rate of energy expenditure in a reach of channel are less compatible with conditions of equilibrium than is a more or less continuous or uniform rate of energy loss. It may well be that a meandering channel is most stable when the energy loss due to curvature is at a minimum. Such a conclusion is perhaps implied in the modal distribution of values of the ratio of radius of curvature to width in natural streams.

If we view curvature as simply one method of altering the distribution of energy expenditure in a given length of channel, it is clear that the pattern of meandering will respond to changes in discharge and load. It is well known that an increase in discharge in a meandering channel will increase the channel width and will increase the size of the meander bends. Decrease in discharge will gradually reverse the process.

In the natural rivers, geologic and stratigraphic evidence clearly demonstrates that during late Pleistocene time an increasing discharge markedly decreased the gradient of Öster-Dal River and enlarged the width of channel as well as the size of meander beds (Wenner and Lannerbro, 1952, p. 108). During this gradual degradation of the valley floor the meandering pattern persisted. Thus the meander pattern where it exists in nature appears to be a persistent attribute of the river.

Change in load will cause aggradation or degradation and thus change in channel slope and size of bend (Friedkin, 1945, p. 7–9); an increase in slope will produce an increase in meander length and amplitude. Schoklitsch (Shulits, 1935, p. 644–646) and Bagnold (1960) have postulated that at high discharges sediment transport is a function of the rate of work done per foot of channel length, or power intensity. In a meandering channel in equilibrium increasing tightness of bend (curvature) through its effect on the rate of energy expenditure decreases the rate of transport. For equilibrium, then, a balance must be maintained between curvature and transport quite apart from any change in intensity of energy loss brought about by a change in length.

Direction for Future Work

No wholly adequate explanation of meandering is yet available. Probably no single simple mechanism will suffice to explain all aspects of meandering. Although there is general agreement on the manner in which bank erosion and point-bar formation are related to the orderly transfer of sediment which is basic to meandering, no physical or mechanical principle has been identified which explains qualitatively the size and geometry of meander curves.

There is need to investigate oscillatory forces which might explain more adequately the manner in which an initial bulge or depression in a stream bank leads to a symmetrical reversal of curvature.

Although the available velocity distributions in channel bends do permit general descriptions of the flow, many characteristics, including orientation and position of the helical patterns, are as yet poorly defined. More detailed measurements in natural channels are required to define the loci of energy losses and their relation to flow resistance and localized erosion of bed and banks.

The way in which a natural channel distributes the energy loss as between boundary friction, form resistance, curvature, and transport is little understood. Without such understanding it is virtually impossible to explain or predict the behavior of a meandering channel. Although some studies of energy losses in curved channels have been made (Allen, 1939; Leopold *et al.*, in press), it appears to us that laboratory as well as field studies on the distribution of energy expenditure in straight and curved alluvial channels are needed. It would be particularly desirable to map the distribution and magnitude of boundary shear in bends of different patterns but with similar cross section and depth. These observations must in turn be related to the mechanics of sediment transport in meandering channels of different patterns.

The principal unsolved problem with regard to the pattern of flow and its relation to erosion and deposition is in the area of the mechanics of sediment transport. Present theories are inadequate to explain the transport of heterogeneous sizes under the variety of conditions found in nature. The stress needed to induce and maintain motion is probably different for scattered rocks on a sand bed, uniform grading from sand to cobbles, graded cobbles and boulders without sand, or a few isolated boulders on a cobble bed. At the same time, objective criteria are needed to describe the ability of varying bank materials to withstand erosive stresses. A quantitative explanation of the meandering process will require balancing the erosive stress produced by tangential shear of the flowing water and the comparable resisting stress provided by the bank material.

It is particularly important that critical field observations be tied to theoretical and laboratory studies. A few fundamental concepts can unify a vast amount of empirical observation if the observations include those critical measurements which make possible an adequate test of theory.

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Appendix. Shapes of Meander Waves in Alluvial Plains

Appendix.—Shapes of Meander Waves in Alluvial Plains

(*Note:* Reaches of river chosen from maps to represent (a) a single meander length in which the S^3 curve was reasonably symmetrical; (b) the samples would include a large range in sizes of rivers. Each sample is considered representative of the local reach of river.)

River and location	Elevation (ft)	Range in width (ft)	Mean width (W) (ft)	Meander length (λ) (ft)	Arc distance (ft)	Mean radius (r_m) (ft)	Amplitude (A) (ft)	Sinuosity	Map quadrangle (scale = 1:24,000)
Sacramento R nr Chico, Calif.	110	250–550	550*	6700	8400	2050	2150	1.25	Ord Ferry, Calif.
Ninnescah R nr Belle Plain, Kans.	1200	80–160	260*	1350	1700	360	400	1.26	Belle Plain, Kans.
New R nr Brawley, Calif.	150	70–90	250*	1475	1750	210	400	1.18	Brawley, Calif.
Missouri R nr Buckner, Mo.	700	350–1550	1850*	14,600	24,300	3150	8850	1.66	Buckner, Mo.
Missouri R nr Buckner, Mo.	690	550–1950	1600*	37,100	41,800	15,550	7700	1.13	Buckner, Mo. Camden, Mo.
Kansas R nr Eudora	790	220–800	1000*	12,750	14,400	4300	2850	1.12	Lawrence E. Kans. Eudora, Kans.
Kissimmee R nr Okeechobee, Fla.	18	85–155	180*	1050	2900	205	1250	2.76	Okeechobee N.W., Fla.
Arkansas R nr Mulvane, Kans.	1190	220–360	500*	5650	7900	1350	2150	1.40	Mulvane, Kans.
Colorado R nr Blythe, Calif.	270	450–1650	1050*	25,300	33,300	7300	8900	1.32	Blythe N.E. Calif. Ariz.
San Joaquin R nr Patterson, Calif.	35	120–280	400*	2400	4550	610	1600	1.90	Brush Lake, Calif.
James R nr Forestburg, S. D.	1220	60–90	160*	2220	3600	450	1275	1.62	Forestburg, S. D.
James R nr Clayton, S. D.	1190	80–150	230*	1550	2600	265	900	1.68	Tschetter, S. D.
Souris R.	1480	30–60	150*	1100	2500	160	720	2.27	Voltaire, N. D.
Missouri R nr Lexington, Mo.	690	450–1150	1350*	24,400	29,000	6000	7100	1.19	Camden, Mo. Lexington, Mo.
Sacramento R nr Glenn, Calif.	88+	250–450	950*	7600	10,800	2300	3200	1.42	Glenn, Calif. Llanos Seco, Calif.
Red R nr Campti, La.	100	400–1200	1400*	11,400	18,700	2130	7000	1.64	(scale 1:62,500) Campti, La.
Henrys Fork nr Menan, Idaho	4810	180–300	340*	2700	6100	850	2300	2.26	Menan Buttes, Idaho
Henrys Fork nr Menan, Idaho	4812	110–200	280*	1950	3300	410	1200	1.69	Menan Buttes, Idaho
Cedar R nr Belgrade, Neb.	1700	60–130	200*	3050	4550	750	1450	1.49	Belgrade, Neb.
Cedar R nr Belgrade, Neb.	1650	90–150	210*	2300	2900	575	820	1.26	Belgrade, Neb.
Mississippi R nr Wynnburg, Tenn.	265±	2132–5460	4940*	38,580	60,050	9620	27,400	1.55	*** Map 6, miles 92–102
Reach		22–48	30	205	310	41	94	1.51	Plane-table map, authors
Squaw Cr nr Lander, Wyo.			8	82	115	17	40	1.40	Plane-table map, authors
Little Pipe Cr nr Westminster, Md.		10–18	12	145	192	35	58	1.32	Plane-table map, authors
Baldwin Cr nr Lander, Wyo.		10–18	14	185	245	45	72	1.32	Plane-table map, authors
Little Sandy nr Elkhorn, Wyo.		10–16	13	248	273	76	55	1.10	Plane-table map, authors

Reach		22–48	30	205	310	41	94	1.51	Plane-table map, authors
Squaw Cr nr Lander, Wyo.			8	82	115	17	40	1.40	Plane-table map, authors
Little Pipe Cr nr Westminster, Md.		10–18	12	145	192	35	58	1.32	Plane-table map, authors
Baldwin Cr nr Lander, Wyo.		10–18	14	185	245	45	72	1.32	Plane-table map, authors
Little Sandy nr Elkhorn, Wyo.		10–16	13	248	273	76	55	1.10	Plane-table map, authors
Buffalo Fork at Black Ranger Sta.		70–115	80	990	1230	308	310	1.24	Plane-table map, authors
Baldwin Cr nr Lander, Wyo.		12–24	15	165	205	43	48	1.24	Plane-table map
									Leopold and Miller
Mississippi R nr Smithland, La.	15	4900–2180	3330	65,000	75,400	15,700	19,850	1.16	*** Maps 38, 38, miles 770–785
Mississippi R nr Lake Providence, La.	85	2200–5800	3500	53,500	75,800	9900	11,400	1.42	*** Maps, 26, 27, miles 550–565
Mississippi R nr Rosedale, Miss.	145	4300–2440	3210	40,300	47,000	11,200	11,760	1.17	*** Maps, 19, 20, miles 387–396
New Fork nr Pinedale, Wyo. (Hailstone Reach)		48–76	62	745	960	163	225	1.29	Plane-table map, authors
Model, U. S. Waterways Expt. Sta.		2–4.8	3.3	34	43.5	8.9	13.0	1.28	Meandering of alluvial rivers, Friedkin, 1945 — Test 4, Plate 41
Model, U. S. Waterways Expt. Sta.		1.2–3.2	1.6	28.2	36.0	7.2	9.5	1.28	Test 1, Plate 29
Model, U. S. Waterways Expt. Sta.		2.3–3.0	2.6	29.5	34.2	7.0	8.5	1.16	Test 3, Plate 37
Moyen R nr Castets, France		627–1120	870	9040	12,200	2000	3760	1.36	Fargue, 1908
Coosa River	440	650–400	520	8976	18,500	1700	6700	2.06	Riverside, Ala.
Kansas River	780	850–170	510	16,400	26,100	3150	8550	1.59	Eudora, Kans.
Red River	100	500–1000	750	10,000	17,500	2000	5000	1.75	Campti, La.
Henrys Fork, nr Menan, Idaho	4810	350–150	250	2700	5500	650	2000	2.04	Menan Buttes, Idaho
Sacramento River	85	600–220	410	6100	11,000	1700	3700	1.80	Glenn and Llano Seco, Calif.
Kissimmee River	18	130–110	120	1100	2100	225	600	1.91	Okeechobee N.W., Fla.
Sacramento River	120	700–350	525	8340	11,500	1400	3600	1.38	Ord Ferry, Calif.
James River	1220		110	2220	2600	600	600	1.17	Forestburg, S. D.
Souris River, N. D.	1610		140	1800	3200	440	1040	1.78	** Sheet A, mile 12, 1926
Souris River	1590		140	1840	4000	480	1240	2.17	** Sheet A, mile 40, 1926
Souris River	1560		120	840	1600	220	500	1.90	** Sheet Cm mile 123, 1926
Souris River	1430		100	1140	2400	300	800	2.10	** Sheet K, mile 290, 1930
Red River, La.	40		600	7400	8800	1640	2100	1.19	*** Mile 33.0

* Estimated bankfull width

** Plan and profile of Souris (Mouse) River, International Boundary to Verendrye, N. D. (Advance sheets): Department of the Interior, Geological Survey, printed 1926

*** Mississippi River Commission, 1935, Maps of the Mississippi River, Cairo, Ill., to the Gulf of Mexico, La.; scale, 1:62,500: Vicksburg

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ASH FLOWS

BY ROBERT L. SMITH

ABSTRACT

The ash flow is the basic unit of most pyroclastic deposits known as welded tuffs, tuff flows, pumice flows, or ignimbrites. Other useful units are the cooling unit, both simple and compound, the composite sheet, and the ash-flow field. The deposit of one ash flow is considered analogous to, but not necessarily the same as, the deposit produced by the passage of one nuée ardente. The geological evidence favors gas-emitting particulate flows as the agents of transport.

The ash-flow is an efficient heat-conserving mechanism, and, because many ash-flow fields contain cooling units believed to be derived from common magma, but clearly emplaced at different temperatures, a cooling mechanism must be invoked. Such a mechanism may be the vertical eruption column which influences the final properties of many ash flows.

Deposits are grouped in seven orders of magnitude ranging from 0.001 to 10,000 km³. Orders 1 to 3 include deposits erupted from domes. Those from craters rarely exceed 10 km³, order 4. Deposits of orders 5 to 7 are associated with subsidence structures in all examples where the source area is known. Welded tuffs occur in fields of all volumes, but they are common among the fissure-erupted deposits of larger volume. Some ash-flow fields distributed from a common source area reach areal dimensions of more than 12,000 square miles and volumes of about 500 cubic miles; some multiple-source fields are known to have volumes of more than 2000 cubic miles. Deposits of single-eruption cycles having volumes of more than a few cubic miles are thought to be related to subsidence structures.

Sorting data suggest that most welded tuffs contain more than 70 per cent by weight of materials less than 4 mm in diameter. Welding and crystallization depend largely on relative temperature and thickness of the cooling units, and most of the textures in welded ash flows can be explained in terms of these variables. Welding begins with incipient cohesion of glass shards and fragments and continues with decreasing pore space and deformation to complete welding which results in a dense black glass.

Crystallization is superimposed on the welded material at any stage, but the degree of welding may influence the type of crystallization. The degrees of welding and of crystallization are zonal and permit the distinction between simple and compound cooling. Incipient welding may take place below 535° C., but complete welding depends on load pressure and time and is thus influenced by the cooling history of individual cooling units. Deposits emplaced at low temperatures such as the Crater Lake, Oregon, "pumice flows" are essentially nonwelded in thick deposits. The high-temperature welded tuffs of southeastern Idaho are densely welded in very thin sheets. Most welded tuffs are probably emplaced at temperatures intermediate between these limits. Some deposits record a systematic change in temperature during the eruption cycle; others record a change in magmatic composition. Some ash flows are thought to be genetically related to near-surface plutons, some of which are large and complex.

RÉSUMÉ

La coulée de cendres est l'élement de base de la plupart des dépôts pyroclastiques appelés tufs soudés, coulées de tufs, coulées de ponce, ou ignimbrites. D'autres élements utiles sont le bloc de refroidissement, simple ou composé, la nappe composée, et le champ de cendres de coulées. Les dépôts d'une coulée de cendres sont considérés comme analogues, mais pas nécessairement identiques aux dépôts produits par le passage d'une seule nuée ardente. Les données géologiques sont en faveur de coulées en menues particules, et émettant des gaz, comme agents de transport.

La coulée de cendres est un mécanisme effectif de conservation de la chaleur, et parce que beaucoup de champs de cendres contiennent des blocs en refroidissement dont on pense qu'ils dérivent d'un magma commun, mais apparemment mis en place à des températures différentes, un mécanisme de refroidissement doit entrer en jeu. Un mécanisme

de ce genre pourrait être la colonne d'éruption verticale qui influe sur les propriétés ultimes de nombreuses coulées de cendres.

Les dépôts sont groupés en 7 ordres de grandeur, de 0,001 à 10.000 km³. Les ordres de 1 à 3 comprennent les dépôts venus de dômes. Ceux originaires de cratères dépassent rarement 10 km³, ordre 4. Les dépôts des ordres 5 à 7 sont associés à des structures de subsidence dans tous les exemples dont le lieu d'origine est connu. Les tufs soudés forment des champs de tous volumes, mais ils sont communs dans les dépôts d'origine fissurale et de grand volume. Certains champs de cendres venus d'un lieu d'origine commun atteignent des surfaces de plus de 30.000 km² et des volumes de plus de 2.000 km³; des champs d'origine multiples atteignent des volumes de 8.000 km³. On pense que les dépôts de cycles à une seule éruption qui ont des volumes de plus de quelques km³ sont en relation avec des structures de subsidence.

Les données granulométrique suggèrent que la plupart des tufs soudés contiennent plus de 70% en poids de matériaux de moins de 4 mm de diamètre. L'agglomération et la crystallisation dépendent largement de la température relative et de l'épaisseur des unités en refroidissement, et la plupart des textures rencontrées dans les coulées de cendres agglomérées peuvent d'expliquer en considérant ces variables. L'agglomération commence avec la cohésion naissante d'éclats et de fragments de verre, et continue par la diminution du volume des pores, et déformation, jusqu'à soudure complète qui aboutit à un verre noir dense.

La crystallisation se surajoute à n'importe quel stade dans le matériel aggloméré, mais le degré de soudure peut influer sur le type de crystallisation. Les stades d'agglomération et de crystallisation sont en zones, et permettent de distinguer le refroidissement simple du composé. L'agglomération naissante peut avoir lieu en dessous de 535° C., mais la soudure complète dépend de la pression et de la durée, donc est influencée par l'histoire du refroidissement des blocs individuels en refroidissement.

Les matériaux mis en place à basses températures, tels que les "coulées de ponce" de Crater Lake dans l'Oregon sont en grande partie non soudés dans les dépôts épais. Les tufs soudés à haute température du sud-est de l'Idaho sont densément soudés en couches très minces. La plupart des tufs soudés sont probablement mis en place à des températures intermédiaires entre ces extrêmes. Certains dépôts montrent un changement systématique de température pendant le cycle de l'éruption, d'autres montrent un changement de composition magmatique. On pense que certaines coulées de cendres sont liées génétiquement à des plutons proches de la surface dont certains sont de grande taille et de nature complexe.

Zusammenfassung

Der Aschenstrom ist die Grundeinheit der meisten pyroklastischen Ablagerungen, die als zusammen geschweisste Tuffe ("Schmelztuff"), Tuffströme, Bimsteinströme oder Ignimbrite bekannt geworden sind. Andere anwendbare Einheiten sind die einfache und die zusammengesetzte Abkühlungseinheit, die komplexe tafelförmige Masse, und das Aschenstromfeld. Man nimmt an, dass die Ablagerung eines Aschenstroms im allgemeinen analog zu der ist, welche durch die Passage eines nuée ardente ("Glutwolke") entsteht, aber nicht unbedingt dieselbe sein muss. Die geologischen Gegebenheiten sprechen dafür, dass gasausströmende Partikel-Ströme als Transportkräfte dienen.

Der Aschenstrom ist ein wirkungsvoller, Wärme konservierender Mechanismus. Da viele Aschenstromfelder Abkühlungseinheiten enthalten die wahrscheinlich von gewöhnlichem Magma abstammen, jedoch eindeutig bei anderen Temperaturen eingebettet worden sind nimmt man an dass ein Abkühlungsmechanismus dabei zur Hilfe gekommen sein muss. Die vertikale Eruptionssäule könnte solch ein Mechanismus sein, welcher die endgültigen Eigenschaften vieler Aschenregen beeinflusst.

Die Ablagerungen werden in sieben Grössenordnungen eingeteilt, welche von 0,001 km³ bis 10.000 km³ rangieren. Die Grössenordnungen 1 bis 3 schliessen Ablagerungen ein, die von Domen eruptiert sind. Kraterablagerungen übersteigen selten die Grössenordnung 4 (10 km³). Ablagerungen der Grössenordnungen 5 bis 7 sind in allen Fällen, wo das Ursprungsgebiet bekannt ist, mit Bodensenkungsstrukturen verbunden. Zusammen geschweisste Tuffe kommen in Feldern aller Grössenordnungen vor. Sie sind jedoch unter den aus Spalten eruptierten Ablagerungen grösseren Umfangs häufiger. Einige Aschenstromfelder, die von einem gemeinsamen Ursprungsgebiet aus verbreitet wurden, erreichen Flächendimensionen von über 30.000 km² und Volumen von etwa 2.000 km³. Es ist bekannt, dass einige Felder mit mehreren Ursprungsgebieten Volumen von über 8.000 km³ haben. Es wird angenommen, dass Ablagerungen von Einzel-Eruptions-

cyclen, welche ein Volumen mehr als einigen km³ haben, zu Absenkungs-Strukturen in Beziehung stehen.

Korngrössen-Analysen lassen vermuten, dass mehr als 70% des Gewichtes der meisten zusammen geschweissten Tuffe aus Materialien bestehen, die weniger als 4 mm im Durchmesser sind. Das Schweissen und Auskristallisieren hängt weitgehend von der relativen Temperatur und der Dicke der abkühlenden Einheiten ab, und die meisten Texturen in den zusammen geschweissten Aschenströmen können mit Hilfe dieser Variablen erklärt werden. Der Schweissprozess beginnt mit geringer Kohäsion von Glasscherben und -bruchstücken und verläuft weiter über abnehmenden Porenraum und Deformationen bis zu einem völligen Verschweissen, das in einem undurchsichtigen, schwarzen Glas endet.

Kristallisation kann in dem geschweissten Material in jedem Stadium auftreten, aber der Grad des Zusammenschweissens kann die Art der Kristallisation beeinflussen. Die stufen von Zusammenschweissung und Kristallisation sind zonal angeordnet und erlauben die Unterscheidung von einfacher und komplexer Abkühlung. Beginnende Zusammenschweissung kann unter 535° C. stattfinden, aber vollständige Zusammenschweissung hängt vom Gewichtsdruck und von der Zeitspanne ab und ist so durch den Abkühlungsverlauf der individuellen Abkühlungseinheiten beeinflusst.

Ablagerungen, welche bei niedrigen Temperaturen eingebettet wurden, wie die "Bimsteinströme" am Crater Lake, Oregon, sind in dicken Lagern ausgesprochen ungeschweisst. Die Tuffe vom südöstlichen Idaho, welche bei hohen Temperaturen zusammen geschweisst wurden, sind in undurchsichtigen und sehr dünnen Lagen zusammen geschweisst. Die meisten zusammen geschweissten Tuffe sind wahrscheinlich bei Temperaturen, die dazwischen liegen, eingebettet worden. Einige Ablagerungen weisen einen systematischen Temperaturwechsel während des Eruptionsreihes auf, andere zeigen einen Wechsel in der magmatischen Zusammensetzung. Man nimmt an, dass einige Aschenströme genetisch mit nahe der Erdoberfläche liegenden Plutonen verwandt sind, von denen einige von grossem Ausmass und komplexer Natur sind.

ПОТОКИ ВУЛКАНИЧЕСКОГО ПЕПЛА

Роберт Л. Смит

Резюме

Поток вулканического пепла является основной составной частью большинства пирокластических отложений. Эти отложения представлены сваренными туфами, туфовыми потоками, потоками пемзы, или игнимбритами. Другими элементами пирокластических отложений являются простые и сложные залежи, образовавшиеся в результате остывания, сложные покровы и отложения раскаленных лавин пепла. Отложение потока пепла аналогично отложению образованному после прохождения раскаленного облака. Однако это не означает, что эти отложения тождественны. Геологические данные говорят о том, что потоки мелких частиц, выделяющие газ, играют значительную роль в переносе вулканического материала.

Поток вулканического пепла хорошо сохраняет тепло. Процесс остывания играет большую роль в образовании отложений потоков, содержащих остывшие залежи. Эти залежи, очевидно, связаны с общим магматическим источником и образовались при разной температуре. Процесс остывания может начаться с образования вертикального столба изверженного материала, который влияет на конечные свойства потоков.

Отложения потоков подразделяются по размерам на семь групп и их объем колеблется от 0,001 до 10 000 кубических километров. Группы от 1 до 3 включают извержения вулканических куполов. Отложения 4-й группы представлены продуктами извержения кратеров и редко превышают объем 10 кубических километров. Отложения групп от 5 до 7-й ассоциируются со структурами оседания во всех случаях, когла известен источник изверженного материала. Сваренные туфы встречаются в отложениях всех объемов, но они обычны среди отложений трещинных извержений большого объема. Некоторые отложения потоков вулканического пепла, происходящие из одного источника, покрывают площадь более 30 000 квадратных километров и достигают объема около 2 000 кубических километров. Некоторые отложения, имеюще несколько источников, дос-

тигают объема более 8 000 кубических километров. Отложения одиночных циклов извержений, имеющие объем равный нескольким кубическим километрам, вероятно связаны со структурами оседания.

По данным гранулометрического анализа большинство сваренных туфов состоит на более чем 70% по весу из материала с диаметром зерен менее 4 мм. Спайка и кристаллизация зависит, в основном, от относительной температуры и толщины остывающих залежей. Большинство структур сваренных потоков вулканического пепла зависит от температуры и толщины залежи. Спайка начинается со сцепления осколков вулканического стекла и прододжается попутно с уменьшением объема пор и с деформацией составных частей, вплоть до полного сваривания, в результате которого образуется плотное черное стекло.

Кристаллизация накладывается на сваренный материал в любой стадии развития, но степень спайки может влиять на тип кристаллизации. Степень сварки и кристаллизация имеют зональный характер и позволяют различать простой и сложный процесс остывания. Начальная спайка может произойти при температуре ниже 535° по Цельсию, но полная сварка зависит от давления вышележащего материала и от времени, и таким образом зависит от процесса остывания отдельных залежей.

Отложения, образовавшиеся при низкой температуре, такие как "потоки пемзы" Кратерного Озера в Орегоне, в основном, в залежах большой мощности не сварены. Высокотемпературные сваренные туфы из юговосточного Айдахо плотно сварены в очень тонких покровах. Большинство сваренных туфов, вероятно, отложено при средних температурах. Некоторые отложения указывают на систематическое изменение температуры во время цикла извержения, другие указывают на изменения в составе магмы. Существует мнение, что некоторые потоки вулканического пепла генетически связаны с близповерхностными плутонами, некоторые из которых имеют значительные размеры и сложный состав.

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Introduction

The evolution of thought surrounding any facet of scientific knowledge is a fascinating subject, and none more so than that leading to our present concepts of the recognition, origin, and mode of eruption of the rocks commonly called welded tuffs or ignimbrites. Many keen geologists have been troubled or even confused by these rocks because they show characteristics of both tuffs and lavas, but are not clearly assignable to either category.

To this group of rocks belong the "tufflavas" of Russian Armenia, the "piperno" of Italy, the "aso lavas" and "hai ishi" (ash stones) of Japan, some of the "rhomb porphyry" of Norway and Sweden, and perhaps the eutaxites of the island of Teneriffe. We are also finding that tens of thousands of square miles of volcanic rocks, formerly considered lava flows, are pyroclastic rocks firmly indurated by welding or crystallization of their glassy parts.

These rocks are predominantly silicic—rhyolites, quartz latites, rhyodacites, and dacites. Andesites and trachytes may occur in this form, but basalts of this nature are exceedingly rare. As reflected by their composition, these rocks are, for the most part, distributed along the orogenic belts, and they occur throughout the geologic column.

With few exceptions, the rocks in question are emplaced as particulate flows or pyroclastic avalanches in contrast to fluidal lava flows and air-fall pyroclastic rocks. The eruptive mechanism is generally held to be the nuée-ardente or glowing-cloud eruption in one or more of its many forms, although some authors have taken exception to the nuée-ardente eruption as the universal agent, especially for welded tuff sheets and "pumice flows" of very large volume.

The study of welded tuffs and of their nonwelded counterparts is still in its infancy, although major progress is being made, and more can be expected as quantitative results from detailed studies become available.

The importance of these rocks is not limited to pure petrography, petrology, or volcanology. It concerns the mapping of certain volcanic and metamorphic terrains, with their problems of stratigraphic and structural correlation. It seriously affects the study of calderas, volcano-tectonic depressions, batholiths, and many ore deposits. The geochemist should find welded ash flows a fertile field for innumerable problems of magmatic volatiles, trace elements, and rock alteration. To the experimental geologist they offer unique materials for the study of rate processes, such as welding and deformation of glass, devitrification, diffusion, and cooling.

It would be presumptuous to attempt discussion of all the facets of the welded-tuff problem; some have recently been or will be treated in other papers, and on others little or no work has been done. In this review the writer attempts to outline some of the more promising avenues for future research, as well as to concentrate on those facets of the problem that he feels need special consideration.

The reference list contains only a selection of the available papers on the subject. Many Russian and Japanese publications which might have been included were not available in a language the author reads, and other very good papers were simply not sufficiently germane to the subjects selected for discussion. Most of these are listed in the recently published and nearly complete *Ignimbrite bibliography* by Cook (1959). In addition an excellent summary of eruptive processes and their products has been published by Williams (1957, p. 59–65); a very good summary paper covering many facets of the subject has been published by Martin (1959); and Ross and Smith (in press) present a historical review, summarize criteria for recognition, and discuss microscopic and field features and processes.

Acknowledgments

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also due P. E. Cloud, Jr., S. S. Goldich, and J. H. Mackin for critical review of the manuscript.

Nomenclature

The writer recognizes that the nomenclature of these rocks is controversial. Differences on terminology are usually resolved when those most concerned reach a common level of knowledge, mutual understanding, and communication. With these thoughts in mind and with the additional thought that the "welded-tuff problem" is, in many respects, still an "infant," it does not seem profitable to discuss the relative merits of different nomenclatures. Instead, for the demands of internal consistency in this paper, the writer will present briefly the usage he has found most applicable to the greater number of deposits of welded tuffs or their nonwelded equivalents. He has no illusions that this usage will necessarily satisfy the needs and beliefs of other workers, nor does he wish to create the impression that it should.

ASH FLOW: The basic unit of ash-flow deposits is the ash flow that is analogous to, or perhaps the same as, the deposit resulting from the passage of one nuée ardente. As defined here, it consists of 50 or more weight per cent of ash and fine ash exclusive of foreign inclusions. Ash (¼ to 4 mm) and fine ash (less than ¼ mm) refer to the Wentworth and Williams' (1932, p. 45) size scale. Most glass shards and crystals are of ash or fine-ash size. The remaining essential materials are pumice lapilli and blocks or, in some types, scoriaceous materials, or mixtures of the two. Most welded tuffs and their nonwelded equivalents fall in this category (*see* discussion of sorting); however, some pyroclastic flows will not conform to the arbitrary 50 per cent minimum for essential particles less than 4 mm in diameter. For these the term lapilli-and-block flows might be used, but as this term is cumbersome, and in view of the fact that the lapilli and blocks can be separated on the basis of their pumiceous or scoriaceous nature, pumice flow or scoria flow (whichever material predominates) seems appropriate, but would call for redefinition of these terms. The term ash flow as used here includes all or, within the limits of the defined particle size, most of the deposits described under the following terms:

Tuff flow
Incandescent tuff flow
Sand flow
Pumice flow
Scoria flow
Block and ash flow (in part)
Gas and ash flow
Hot ash flow
Glowing avalanche (in part)
Deposits of the St. Vincent vertical type (in part)
Deposits of the "Nuée Ardente du katmaiennes"
Deposits of the Valley of Ten Thousand Smokes type
Aso lava
Ignimbrite
"Hai ishi"
Tuff lava
Welded tuff
Welded mud lava
Schmelztuff
Piperno
"Shirasu"
Sillar

ASH-FLOW SHEET: Some ash-flow deposits have not been subdivided into well-defined units, yet they are mappable, sheetlike stratigraphic units and are of inferrable ash-flow origin. For these a nonspecific name is useful. Ash-flow sheet is used in this sense to include any unspecified sheetlike unit or group of units considered to be of ash-flow origin. If the materials of the sheet are predominantly welded tuff or if it is desirable to emphasize their welded character, the term welded tuff sheet is used.

WELDED ASH FLOW: An ash flow in which welding has produced a conspicuous zonal pattern is termed a welded ash flow.

WELDED TUFF: A welded tuff is a rock or rock body in which vitric particles have some degree of cohesion by reason of having been hot and viscous at the time of their emplacement. Incipient, partial, dense, or complete welding are indicated by these terms. The term welded tuff does not carry a stratigraphic connotation. However, a deposit of welded tuff could conceivably include an entire ash flow, and probably does in some cooling units. The term welded tuff should exclude "fused tuffs" in which reheating of vitric pyroclastic beds in proximity to hot rock bodies was responsible for their coherence. Crystallization may follow or accompany welding but, while commonplace, is not essential to that process. Some air-fall materials may be included where welding was in consequence of the inherent heat of the particles.

WELDED TUFF SHEET: A sheetlike body of rock containing conspicuous welded tuff is called a welded tuff sheet. Used loosely it

refers to any unit or group of units that have not been specifically defined.

COOLING UNIT: A single or multiple ash-flow deposit (Pl. 1) that can be shown to have undergone continuous cooling is a cooling unit. The cooling history of the unit may have been essentially uniform or broken by successive emplacements, but not broken to the extent that one part cooled completely before another part was emplaced on top of it.

A Simple Cooling unit is an ash flow or sequence of ash flows that has had an essentially uninterrupted cooling history. Such units remain either nonwelded and noncrystalline during cooling or form a pattern of textural zones. The zonal pattern will be consistent with the individual welding and crystallization characteristics of the unit, as determined by the variables emplacement temperature, thickness, gas content, and physical and chemical composition. (*See* section on Zones and Zonal Variations.)

A Compound Cooling unit is one that shows departures in the expectable zonation and other properties which result from simple cooling, because the intervals between ash flows were too great for readjustment to a single-unit cooling gradient (Pl. 2, fig. 1).

COMPOSITE SHEET: A cooling-unit complex that grades from one cooling unit into two or more cooling units is a composite sheet.

ASH-FLOW FIELD: Deposits of pyroclastic rocks consisting preponderantly of ash flows, which are related to some specific unit of area, constitute an ash-flow field.

Ash-Flow Problem

Within the many-faceted ash-flow problem the writer stresses those deposits that contain welded tuffs, especially those that occur in great volume over large areas. The solution of the many unsolved problems concerning these rocks will come about only through detailed studies. Some of these problems are minor ones, but at least five must be considered major: (1) recognition of ash flows, (2) recognition and definition of units, (3) mode of eruption, (4) eruptive vents and source areas, and (5) origin of the magmas.

Twenty-five years ago Marshall (1935) and Mansfield and Ross (1935) discussed the field and microscopic characteristics of the New Zealand "ignimbrites" and the Idaho "welded tuffs." Since that time nearly all authors who have studied welded tuffs in detail have discussed problems of recognition, and many have suggested criteria. The problem still remains, however, because the distinction between welded tuffs and lava flows on the one hand, and between ash-flow tuffs and tuffs of other origins on the other hand, is not always possible.

Recognition and definition of units within ash-flow fields is especially important where detailed structural and stratigraphic correlations are sought and where genetic problems are concerned. Deposits that were emplaced virtually "en masse" or from a very rapid sequence of eruptions are much less complex than those that contain cooling breaks or erosional disconformities.

Problems of the mode of eruption are related to the fact that no scientifically observed eruptions are known to have produced welded tuffs or deposits comparable in volume to those resulting from many prehistoric eruptions. Nearly all reasoning of the last 35 years concerning prehistoric deposits has been by analogy with the Valley of Ten Thousand Smokes whose deposits, in turn, have been explained in terms of the nuée-ardente concept developed from the observed eruptions of Pelée and La Soufrière.

The soundness of this chain of reasoning is rarely questioned, although some misinterpretation of the classical concepts have been published. Nevertheless, the eruption mechanisms before emplacement of deposits from "nuées ardentes" need investigation and clarification as does the precise meaning for the term nuée ardente, if we persist in using the term to designate the agent of emplacement of these deposits.

The most important ash-flow problem concerns the location and kind of eruptive vents and source areas which produce the sheet deposits of great volume. It seems almost incredible that, of the hundreds of occurrences of ash flows comprising tens of thousands of cubic miles of rock in the western United States, for example, the number of unequivocally known source areas is probably less than 10. In terms of published data the number is 4. These are all subsidence structures: Crater Lake caldera, Oregon; Valles caldera, New Mexico; Creede caldera, Colorado; and parts of Yellowstone National Park. On the other hand, practically all the deposits of "welded tuffs" or "pumice flows" known to have a volume of more than a few cubic miles and that have been unequivocally related to their source areas are related to subsidence structures. With this evidence favoring subsidence in the source

areas, it becomes of prime importance to locate these sources for the older deposits.

The question of origin of magma is of special interest in pyroclastic eruptions producing ash flows, because some ash-flow deposits are very large and are located in tectonic settings that suggest kinship with granitic batholiths. The problem concerns the immediate origin of the magma from "volcanic" or "plutonic" sources rather than its ultimate origin.

Mode of Eruption

General Statement

Present concepts of the mode of eruption of hot pyroclastic flows center around the nuée-ardente or glowing-avalanche eruption and depend almost entirely on the well-known published accounts of the 1902 eruptions of Mt. Pelée in Martinique (Lacroix, 1903a; 1903b; 1904; Anderson and Flett, 1903) and La Soufrière on St. Vincent (Anderson and Flett, 1903), and the eruption in 1912 near Mt. Katmai in the Valley of Ten Thousand Smokes, Alaska (Fenner, 1923). These accounts have been augmented by detailed observations of the 1929–1932 eruption of Pelée by Perret (1935) and by various eruptions of Merapi (1930–1935) in central Java (Stehn, 1936; Neumann van Padang, 1933; Van Bemmelen 1949). More recent eruptions, among which were those from Lamington (1952) in New Guinea (Fisher, 1954; Taylor, 1954; 1956) and Hibok-Hibok (1948–1953) in the Philippines (MacDonald and Alcaraz, 1956), have tended to confirm and strengthen the correctness of earlier observations.

There is thus a body of data from several competent observers which makes the "nuée ardente," in at least some of its forms, well known.

Opinion is divided, however, as to whether or not nuée-ardente mechanics are adequate to explain the eruption and emplacement of the pyroclastic materials that sometimes form welded tuffs. It may be significant, in this regard, that no observed nuée ardente has formed a recognized welded tuff. Because of the divided opinion and because some of the existing concepts of "welded-tuff" emplacement seem traceable to misconceptions regarding the classical nuées ardentes and to ambiguous usage of the term, it seems necessary to inquire briefly into the nature of the classical nuées ardentes. This is also important for understanding the development of the "pyroclastic-flow" concept.

Nuée Ardente

Lacroix (1903a) introduced nuée ardente as a geological term for the devastating eruption of Mt. Pelée on May 8, 1902, and for subsequent eruptions from Pelée. The eruption on May 7, 1902, of La Soufrière on St. Vincent also came to bear the name nuée ardente, as have many other eruptions from other volcanoes in other parts of the world, particularly Merapi in central Java.

Nuée ardente has also been used to designate eruptions which are believed to have been responsible for the great welded-tuff sheets and their nonwelded counterparts which are the main subject of this paper. This usage of nuée ardente has its origin in the studies of Fenner (1923), Lacroix (1930), and others, although Fenner sensed the fundamental differences between the deposits of the Valley of Ten Thousand Smokes and those of Pelée.

It should now be recognized that most welded ash flows cannot have had an origin from an eruption of the type that occurred at Mt. Pelée in 1902. The very size of the deposits precludes this; however, the mechanics of eruption and transport as deduced by the observers of the deposits of the eruption of Pelée and La Soufrière that took place on May 7 and 8, 1902, and observers of subsequent eruptions from Pelée, together with Fenner's interpretation of the deposits of the Valley of Ten Thousand Smokes, give us insight into a probable mechanism for eruption and transport of the ash-flow deposits of large volume.

All observers of the nuée ardente phenomenon except possibly Lacroix, agree that the nuée ardente has two parts—a basal avalanche and an overriding spectacularly expanding cloud of gas and dust (Lacroix, 1904, p. 350; Anderson and Flett, 1903, p. 447–453; Perret, 1935, p. 5; Neumann van Padang, 1933, p. 134; Van Bemmelen, 1949, p. 191–192; Taylor, 1954, p. 86; Akkersdijk, 1941, p. 63–64). The distinction between the two parts has not always been sharply drawn. These statements refer to the nuée ardente as an agent of transport of material from the vicinity of the vent to its final resting place, not to the material as it is being erupted from the vent.

Lacroix (1904, p. 350) has described the nuée ardente:

"Une nuée ardente est constituée par une émulsion de matériaux solides dans un mélange de vapeur d'eau et de gaz à haute température, sur la nature desquels je ne reviendrai pas. La forme et les dimensions qu'elle présente au moment de sa sortie de la carapace du dôme montrent qu'elle était soumise sous celle-ci à une compression formidable, car, au bout de quelques secondes, elle occupe un volume supérieur de plusieurs milliers de fois à celui qu'elle possédait lors de son départ . . .

" . . . à la base, se trouve une zone à plus haute température, dans laquelle prédominent des matériaux solides (blocs de toutes dimensions, fragments plus petits, cendres fines); chacune des parties ou des particules solides qui les constituent rayonne de la chaleur et doit être entourée par une atmosphère de gaz et de vapeurs extrêmement comprimée au début, mais se dilatant rapidement; c'est cette atmosphère qui, empêchant les particules solides de se toucher, maintient l'ensemble dans un état de mobilité lui permettant de couler sur les pentes, presque à la façon d'un liquide."

Probably because of the trajectoral (directed-blast) nature of the May 8 nuée ardente, at least in its early stages of formation, Lacroix was aware of but did not emphasize the basal avalanche; he was, rather, preoccupied with the "blast" or "cloud." Herein seems to lie the roots of the "airborne-cloud" connotations which permeate the literature on welded tuffs or ignimbrites. Many writers have referred to the clouds as the agents of transport of "avalanche" materials. This is doubly confusing when it is also realized that the cloud portion of the nuées ardentes does carry and deposit ash and dust, although in insignificant amounts as compared to the avalanche portion.

Anderson and Flett (1903, p. 494–495), not having seen the trajectoral-type nuée ardente, give the following account of a July 9, 1902, eruption of Pelée:

"Then in an instant a red-hot avalanche rose from the cleft in the hillside, and poured over the mountain slopes right down to the sea. It was dull red, and in it were brighter streaks, which we thought were large stones, as they seemed to give off tails of yellow sparks. They bowled along, apparently rebounding when they struck the surface of the ground, but never rising high in the air. The main mass of the avalanche was a darker red, and its surface was billowy like a cascade in a mountain brook. Its velocity was tremendous. The mist and steam on the mountain top did not allow us to see very clearly how the fiery avalanche arose, but we had a perfect view of its course over the lower flanks of the hill, and its glowing undulating surface was clearly seen. Its similarity to an Alpine snow avalanche was complete in all respects, except the temperature of the respective masses. The red glow faded in a minute or two, and in its place we now saw, rushing forward over the sea, a great rounded, boiling cloud, black, and filled with lightnings. It came straight out of the avalanche, of which it was clearly only the lighter and cooler surface, and as it advanced it visibly swelled, getting larger and larger every minute."

Of the avalanche deposits formed from the May 7 eruption from Soufrière, Anderson and Flett (1903, p. 449) write:

"The geological evidence was not sufficient to demonstrate what form the discharge took, what was its path, and how great was its velocity when it left the crater; but by the time it reached the valleys below it was a rushing torrent of sand, stones, and hot gases, which coursed along the valley bottoms, adapting itself readily to all changes in their configuration, too heavy to surmount any great height, but sweeping over the surface of the minor ridges, and coming gradually to rest in the deep ravines behind them. In the valleys of the Wallibu and Rabaka Dry River it filled the whole channel, and when its energy was spent it lay in banks, with irregular rounded upper surfaces, like glaciers of black sand. In the more shallow and open valley of the Wallibu Dry River it had not been confined to a narrow and steep-sided channel, but had been free to spread out laterally, and there it took the form of a broad sheet of sand with round-backed ridges, like wreaths of snow, pointing down the valley and diverging with a slightly fan-shaped arrangement."

These observations give a very clear picture of the avalanche portion of the nuée ardente. After comparing their observations at Soufrière with those at Pelée, they summarized in part that Pelean-type eruptions are characterized by avalanches of incandescent sand accompanied by a great cloud of gas and dust. The avalanche and the cloud both move at high velocity.

Anderson and Flett believed that the nuées ardentes moved under the influence of gravity, while Lacroix, although recognizing that gravity played a part, believed that a directed initial explosion was also necessary. They agreed that the extreme mobility of the mass was due to expanding gases.

The controversy was slowly resolved when, through observations of other nuée-ardente eruptions, it was realized that the position of the vent within the volcanic edifice influenced the initial behavior of the nuée ardente. It is not necessary to review the details of this evolution here; it is revealed in the works of Perret (1935), Lacroix (1908; 1930), Escher (1933a; 1933b), Stehn (1936), Grandjean (1931), Neumann van Padang (1933), Kemmerling (1932), and MacGregor (1952; 1955). MacGregor's paper is really the first detailed analysis of the problem, and in it many distinctions are clearly drawn. In particular it points up a number of common misconceptions as well as specific published errors.

Fenner (1923, p. 71–74), from his studies in the Valley of Ten Thousand Smokes, concluded that neither gravity nor violent initial explosion, alone, could explain the distribution of the "sand flow," and that a third factor was essential. He writes (p. 74):

"This, as I conceive it, is the self-explosive property inherent in the magma itself, due to the presence and evolution of dissolved gases; and this explosive evolution is not manifested immediately, but is prolonged over a considerable portion of the time during which the spreading movement continues, and gives to the material its fundamental characteristics of quasi-liquidity. Gravity plays an important part in directing the flow down mountain slopes or along topographic depressions, but it is believed that its effectiveness would be rather insignificant were it not for the peculiar properties imparted by the form of disintegration described."

We thus see Fenner as the originator of the "auto-explosive" flow concept. Lacroix, and Anderson and Flett recognized the fundamental importance of expanding gases in giving the mass its high mobility, especially over low gradients, but the delayed or continued evolution of gas has been more clearly expressed by Fenner.

Perret (1935), who witnessed more than 100 nuées ardentes from Pelée between 1929 and 1932, clearly recognized their dual character. His concept of the flowage mechanism was not essentially different from Fenner's. He stressed their wide range in intensity. Van Bemmelen (1949, p. 191–192) summarizes the nuée-ardente phenomenon based on his and other workers' observations in Indonesia, principally from the Merapi volcano in Java. He points out that the character of the eruption changes with an increase in volume of erupted material. Small-volume avalanches may be accompanied by a dust cloud, whereas larger volumes of more than a thousand, up to many hundreds of thousands of cubic meters, give rise to a nuée ardente of dual character. The "ladu" (avalanche) follows the canyons, whereas the "nuée ardente" may jump the banks and become independent of the avalanche for short distances.

The usage of the term nuée ardente is ambiguous. Within the references cited in this section it has been used for (1) the eruption which produces the laterally moving mass of fragmental material and gas, (2) the avalanche and overriding cloud as agents of transport, and (3) the cloud, exclusive of the avalanche.

Seemingly the nuée ardente as an agent of transport of volcanic materials, whatever it may be at the moment of its emergence from the vent, is, when it begins to leave a recognizable deposit, a flow of pyroclastic materials in which the fragments are hot and are emitting gas. An accompanying expanding cloud of gas emerges from the flow and carries considerable solid material, and it may become separated from the flow because of topographic diversions and run a short distance (several miles) on its own, before the gas becomes separated from the solid particles. The cloud may deposit thin beds of ash.

Any argument about the unity or separateness of the avalanche, or flow, and the cloud must center around the degree of inflation of the flow or the sharpness of transition between the flow and the cloud. Some writers have envisioned the entire nuée ardente as a "cloud" with gas suspending the solid materials, the whole having a dense base grading up to a diffuse top. Such a medium should, in spite of turbulence, leave a graded deposit either vertically, laterally, or both. To the writer's knowledge no such deposits have been described.

The eruptions discussed in this section were insignificantly small compared to ancient ash-flow eruptions. To reconcile the two we must magnify the nuées many fold or seek another explanation for the emplacement of ancient deposits.

Perhaps the significant point of difference between Pelean eruptions and eruptions that produce welded ash flows, aside from composition or the character of the vent, is that the entire mass of a Pelean nuée ardente is produced in one short burst, whereas the other type is produced by continuous evolution of material over periods of hours, days, or perhaps weeks. Fluctuations in intensity of the eruption allow diversity in ash flows not seen in Pelean deposits. The difference is probably directly related to the volume of eruptible material.

Vertical Component

The concept of formation of pyroclastic flows from vertical-eruption columns has not been given the consideration it deserves*. To

* Since this manuscript was submitted, two important papers have come to the writer's attention:

Hay, R. L., 1959, Formation of the crystal-rich glowing avalanche deposit of St. Vincent, B. W. I.: Jour. Geology, v. 67, p. 540–562

Taylor, G. A., 1958, The 1951 eruption of Mount Lamington, Papua: Australia Bur. Mineral Resources, Geology and Geophysics Bull. 38, 117 p.

Hay's findings at Soufrière are so pertinent to the discussion of the vertical component and its probable influence on cooling and sorting that the author

be sure, the concept embodied in Lacroix's (1930, p. 421–472) *nuée ardente d'explosion vulcanienne* and the *nuée ardente du massif du Katmai* calls for vertical eruption, but to nearly all writers the vertical component is just sufficient to boost frothing magma to the crater's edge whence it begins its lateral journey as a "nuée ardente." Perhaps the most important of all observed historical eruptions shedding light on the problem was that of Komagatake in 1929. Kozu (1934, p. 136) states that during the early part of the eruption pumice was ejected straight up as high as 12 km. This material was deposited in well-stratified beds; some of it was carried by the winds for more than 200 km. This air-fall pumice was followed by large vertical-eruption columns containing great quantities of pumice erupted to lesser heights than the earlier materials. This later pumice fell back on the mountain and poured down the valleys to distances as great as 6 km. The massive column of pumice above the crater, together with the pumice flows pouring down the mountain, may be seen in one of Kozu's illustrations (Kozu, 1934, Fig. 20).

Williams (1942, p. 79) visualized conditions at Crater Lake, during the eruptions of the "pumice flows," to be much the same as at Komagatake in 1929. Similar deposits were produced, but they are of quite different order of magnitude.

The Komagatake "pumice flows," according to Gilbert (1938, p. 1851), were dry avalanches (avalanches seches, Lacroix, 1930) and are not comparable to deposits such as the Bishop tuff. In one sense this may be true, but it may be argued that, of all pyroclastic flows produced by observed eruptions, the ones most closely analogous to deposits of Crater Lake, Krakatoa, and the Valley of Ten Thousand Smokes are the Komagatake "pumice flows." The Komagatake flows were of very small volume, erupted as a vertical column and hence somewhat cooled by mixing with air before the flows were formed. Yet fumarole temperatures as high as 510° C. were measured in deposits averaging only 10 feet thick.

The Bishop tuff is locally more than 500 feet thick and has a volume of more than 2400 times that of the Komagatake deposits. This alone is probably significant with regard to the ultimate origin, but not with regard to mechanics of emplacement. Units of the Bishop tuff that are only 10 feet thick probably show no high-temperature characteristics that are not shown by the Komagatake flows, and, aside from compositional differences, the two would probably be indistinguishable. This is not to imply that the two deposits were emplaced at the same temperature, but rather that comparisons are not meaningful unless units of equal thickness are found to be different.

That the pyroclastic flow is a heat-conserving mechanism is shown by the temperature measurements on some historic deposits and by the existence of welded tuffs. Boyd (1957, Ph.D. Thesis, Harvard Univ.), in his study of the Yellowstone welded tuffs, has concluded from a careful analysis of factors causing heat loss, both during emplacement of the tuff and in the conduit, that the formation of welded tuff, from material emplaced by the "tuff-flow" mechanism, is thermodynamically feasible.

Boyd's calculations assume minimum mixing with air and indicate a temperature loss of less than 5° C., if the emplacement time was 1 hour, or less than 45° C., if the emplacement time was 10 hours. The approximate correctness of these figures, for the ideal ash flow, is suggested by the existence of certain welded-tuff sheets that are densely welded to within a few feet of their tops. Such a condition calls for very efficient heat conservation, because it means that the glassy particles must deform and weld under a very low load of less than a few pounds per square inch. Temperatures necessary to accomplish welding at such low loads are probably in excess of 800° C. (Smith, Friedman, and Long, 1958), whereas the eruption temperature of rhyolitic magma containing phenocrysts must be below about 960° C. and is probably below 900° C. for most rhyolitic pyroclastic rocks. These figures are suggested from a consideration of Tuttle and Bowen's (1958, p. 58) PT curve for the ternary minimum, and from our knowledge of minimum water contents of volcanic glass (about 0.1 per cent). There is no need to develop this argument here. It is mentioned only to show that field and experimental evidence support Boyd's conclusions that actual heat loss during flowage is small for ideal ash flows and that heat conservation is very efficient.

The ash flow seems in fact to be so efficient a heat-conserving mechanism that the writer is inclined to think that, except for eruptions from domes, the only reason ash flows containing a high proportion of glass ever escape welding is because they were cooled between the time they were erupted from the vent and

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strongly recommends that Hay's paper be read by anyone seriously interested in these facets of the ash-flow problem.

the time they began their journey as pyroclastic flows. Seemingly the only significant cooling mechanism that can be invoked for this interval is mixing with air in a vertical-eruption column; the amount of mixing or degree of cooling is dependent on the height, density, and mechanics of the column, or columns.

Seeking an explanation for Lacroix's "lateral blasts" at Pelée in 1902, Jaggar (1902, p. 366) suggested that they were due to the interference between downfalling ejecta and the upblast from the crater. The interference caused the material to be deflected laterally. Anderson and Flett (1903, p. 513) discounted Jaggar's theory on the grounds that it fails to account for their observations of subsequent eruptions from Pelée. These and later observations by others at Pelée and other volcanoes support the views of Anderson and Flett, and it seems clear that Jaggar's explanation is precluded for eruptions from domes, which, because of the position of their vents, give sectoral distribution and commonly have a horizontal component. However, it is extremely doubtful whether Jaggar's or a similar explanation has been or can be disproved for the 1902 eruption from Soufrière. The important point of Jaggar's explanation is that it demands a vertical column of pyroclastic material and gas.

If the amount of solid material in the column exceeds that which can be efficiently erupted as air-fall pyroclastics, the excess must be displaced laterally as pyroclastic avalanches. Whether this laterally displaced material is influenced more by interference between upsurging and downfalling material, or whether it is related to differential velocities in the column or some other mechanism is an important point, but probably not so important as the thought that the volume of solid material in the column is a function of gas pressure, and that fluctuations or systematic changes in gas pressure might effectively control the amount of, and the emplacement-temperature characteristics of, the laterally displaced material. Such a scheme allows us to envision an eruption column which at one stage is producing air-fall pyroclastics and at another stage is simply boiling fragmental material over the lip of the vent.

Behavior of the laterally displaced material should depend on its volume, temperature, gas content, and the surrounding topography. If its volume and temperature are low, the material may accumulate to form dry avalanches, *avalanches seches* of Lacroix (1930, p. 466) or "hot avalanches" of Perret (1924, p. 90). If its volume is large but its temperature is low, it will perhaps form deposits like those around Crater Lake, Oregon. If its temperature is high and volume variable, it will form welded-tuff sheets, the characteristics of which will be determined by temperature, thickness, gas content, and composition.

The discussion in this section implies that the emplacement temperature of these pyroclastic flows may be controlled to much greater degree by cooling in a vertical column, before the flows form, than by a fundamental difference in magma temperature or initial gas content, and that this cooling is primarily related to the volume of materials in the column, the height of the column, and perhaps the nature of the vent. This is not to infer that the magma temperature is constant; it is not, and some deposits record a systematically changing temperature.

Emplacement by Flowage

The geologic evidence, aside from observed nuées ardentes, for the emplacement of ash flows by flowage or avalanche mechanics has been cited by authors such as Fenner (1923, p. 61–74), Anderson and Flett (1903, p. 449), Williams (1942, p. 81), and Matumoto (1943, p. 3–4). It may be summarized as follows: (1) restriction of deposits to valleys or other topographic lows in mountainous regions; (2) nearly level upper surfaces over buried topography except where compaction because of welding has caused surface irregularities; (3) lack of sorting far from vents which would suggest emplacement *en masse* rather than by air fall; (4) flowage features in some deposits; (5) retention of heat for long periods of time (years) resulting in fumarolic activity, welding, and crystallization, which appear improbable in air-fall deposits; (6) incorporation of surface debris.

TOPOGRAPHIC RESTRICTION. Restriction of ash-flow deposits to topographic lows is well illustrated by the Crater Lake "pumice flows." Williams (1942, p. 81) comments that the outstanding feature of these deposits is the manner in which they have followed the principal canyons near Crater Lake. In the Valles Mountains, New Mexico, all canyons that descend from the high-mountain source area and that are older than the Bandelier rhyolite contain welded ash flows. These flows coalesce to form continuous sheet deposits on pediment and alluvial surfaces below the canyon mouths (Ross and Smith, in press).

In contrast to the coalesced sheets on the lower flanks of the Valles Mountains, the "Aso lavas" (Matumoto, 1943, p. 3–4) form a continuous field peripheral to the caldera, whereas at lower levels the field is divided into innumerable branches that follow the highly dissected topography.

LEVEL SURFACES. The nearly level upper surfaces of ash-flow sheets whose bases may rise and fall in response to changes in topography are well known and need no further discussion.

LACK OF SORTING. Lack of sorting has been stressed as a characteristic feature and criterion for recognition of ash flows. It is generally believed that the lack of sorting is a result of deposition from a turbulent flow or avalanche, although some air-fall deposits in the vicinity of eruptive vents may also give rise to nonsorted tuff deposits, and at times the two are indistinguishable (Williams, 1957, p. 64).

Actually many ash flows do show some sorting, especially near their distal ends, where it can be presumed that they were moving more slowly before coming to rest. Such sorting seems to be related directly to flowage and in particular to some measure of laminar flow, and its visible traces are discussed under primary flow features.

FLOWAGE FEATURES. Flowage features are of two types, primary and secondary.

Primary flow features include:

(1) Pumice swarms—discontinuous streaks or lenticular accumulations of pumice lapilli and blocks in a finer matrix (Taneda, 1954, p. 171; 1957, p. 96; Taneda, Miyachi, and Nishihara, 1957, p. 113).

(2) Rafted pumice—top, side, and flow-front accumulations of pumice blocks (Kuno, 1941, p. 146; Ross and Smith, in press).

(3) Inclusion trains—discontinuous streaks or lenticular accumulations of accidental inclusions (Kuno, 1941, p. 147).

These primary flow features appear to be most abundant near the distal portions of ash flows where they seem to be related to lowered velocity movement before the ash flow comes to rest. Surface accumulations of pumice, marking flow tops, probably are more common than generally recognized and may at times be useful in distinguishing individual flows in densely welded tuff deposits. These flow features may be overlooked in welded tuffs, because welding and crystallization tend to obscure them.

Secondary flow features are:

(1) Radial stretching—shards and particularly pumice fragments may occasionally have been squeezed during compaction and welding beyond simple loss of pore space, resulting in radial stretching of the fragments. Most pumice fragments in densely welded tuffs have simply collapsed without elongation; however, if under great load or high temperature, they occasionally display abnormal elongation in cross section. Ross and Smith (in press) have suggested that secondary solution of volatiles by pumice may cause slight viscosity differences between shards and pumice fragments and may allow preferential radial elongation of pumice by differential compaction. Probably most pumice fragments in ash flows before welding are not longer than three times their smallest dimension. Taking porosity and preferred orientation into account, Ross and Smith (in press) suggest that, after dense welding, collapsed pumice fragments having a vertical dimension 3 to 20 times smaller than their long dimension are probably the result of simple loss of pore space without elongation. An elongation of 20 to 60 or more times the smaller dimension is believed to indicate stretching.

(2) Flowage deformation—the stretched pumice fragments are disclike in plan view and have random orientation. Some welded tuffs, however, contain stretched pumice that has preferred orientation, and these seem to indicate mass flowage during or after welding. In examples of this type seen by the writer the distance flowed could have been only a few inches or at most a few feet.

Secondary flowage in these rocks is of minor importance to the entire problem. It may be related to sliding off topographic highs or earth movements during welding of the tuff.

HEAT RETENTION. Retention of heat, resulting in welding and crystallization, is one of the strongest arguments for the emplacement by flowage of welded tuff deposits. Heat measurements on historic ash-flow deposits are few, but significant. Unfortunately all the temperature measurements on the "sand flow" in the Valley of Ten Thousand Smokes were of necessity made on fumaroles, rather than in the ash itself, and the first of these not until 5 years after the eruption. Fumarole temperatures, however, may approximate the internal heat of the flow, although variables such as gas reactions, localization of gas in joint cracks, depth from point of origin of the gas, and thickness of the flow all make an accurate heat analysis impossible or, at most, very undependable.

Because of the still unresolved controversy over the possibility of a deeper-seated source

for some of the fumaroles, the most significant temperature measurements in the Valley of Ten Thousand Smokes probably are those made on fumarole no. 1, near the distal end of the flow. These are also the only ones that span a 2-year period, and they give some indication of the rate of cooling and of minimum emplacement temperature. They are shown in Figure 1 along with all other temperature measurements made on fumaroles that Allen and Zies could identify in 1919 as the same ones measured in 1918 by Sayre and Hagelbarger (1919, p. 262–278). All but three of these measurements show a temperature drop similar to that shown by fumarole no. 1. The three that show a rise in temperature are unexplained. The maximum temperature measured among the fumaroles was 645° C. (Allen and Zies, 1923, p. 101). If this temperature, 7 years after emplacement of the ash, is unrelated to a deeper source, it certainly must reflect a very deep layer of ash.

Also shown in Figure 1 are cooling curves drawn from measurements of both fumaroles (*S* curves) and the pumice flow deposits (*P* curves) formed during the eruption of Komagatake in 1929 (Kozu, 1934, p. 143). The curves labeled *P* are three of five very similar curves that show the rate of cooling in the pumice at a depth of 70 cm and give a minimum emplacement temperature of 385° C. The curves labeled *S* were made on fumaroles; the maximum temperature of 510° C., measured 8 to 11 days after the eruption, may be close to the maximum emplacement temperatures of the flows.

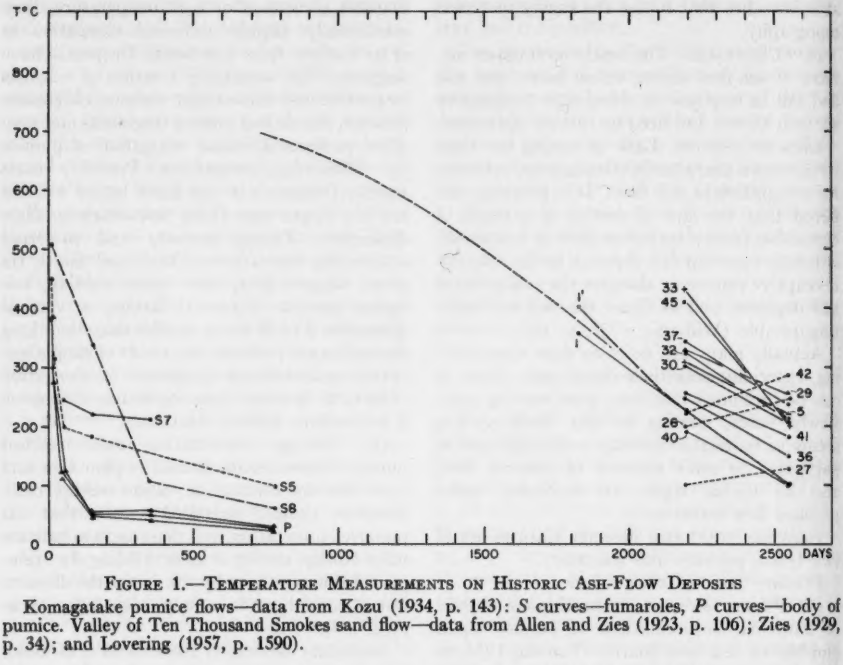

FIGURE 1.—TEMPERATURE MEASUREMENTS ON HISTORIC ASH-FLOW DEPOSITS

Komagatake pumice flows—data from Kozu (1934, p. 143): *S* curves—fumaroles, *P* curves—body of pumice. Valley of Ten Thousand Smokes sand flow—data from Allen and Zies (1923, p. 106); Zies (1929, p. 34); and Lovering (1957, p. 1590).

INCORPORATION OF SURFACE DEBRIS. It is not unusual to see in ash flows inclusions that could have been derived only from the surface over which the ash flow traveled. The inclusions range from very fine-grained material to boulders weighing tons. Some writers have observed evidence suggesting a strong erosive power of ash flows (Tsuya, 1930, p. 252). The ash flows constituting the Bandelier rhyolite tuff of New Mexico have locally swept the underlying surface clear of rock debris of all kinds and sizes up to boulders several feet in diameter, whereas elsewhere the flows have passed over bedded air-fall ash deposits seemingly without disturbing even the finest particles. In those places showing evidence of erosion the determining factor apparently was an irregular ground

surface which increased turbulence in the overriding mass.

Particulate or Foam (Froth) Flows

For at least some "welded tuffs" several writers have proposed mechanisms of emplacement which differ from the pyroclastic-flow concept. It has been suggested that emplacement occurred either as a mass of coherent foam which subsequently collapsed to give "welded-tuff" textures, or as a gas-charged lava which vesiculated and fragmented top and bottom while flowing, resulting in a sort of pseudo "welded-tuff" deposit in which the dense interior of the flow is really a nonfragmented, but vesiculated and collapsed lava.

Kennedy casually (1955, p. 495) suggested the first mechanism for the origin of the vast welded-tuff sheets of the Yellowstone area. Boyd (1957, Ph.D. Thesis, Harvard Univ.) from field and laboratory data, believes the evidence favors the particulate-flow concept; he suggests, however, that one local unit in Yellowstone Park, the Canyon flow, may have originated as proposed by Kennedy.

Hausen (1954) concluded that the Dorena "welded tuff" of western Oregon was emplaced as a vesiculating lava flow. Top and bottom of the flow were fragmented, forming what seems to be typical "pumice tuff," but the interior was never disaggregated. The eutaxitic structure ("welded fluxion structure"), according to Hausen, was formed by "flowage along planes of minute vesicles within the partially expanded lava." Petrofabric study of the biotite flakes in the dacite "welded tuff" (Hausen, 1954, p. 212–213) shows a preferred orientation in the dense interior of the mass, but a less well oriented to random distribution in upper and lower parts.

Hentschel (1955, p. 142) invokes a third mechanism for explaining "welded tuffs." This seemingly involves vesiculation and fragmentation of the magma at some depth without immediate eruption, followed by regeneration ("regenerierte") to a melt, and finally by eruption as nonhomogenized lava flows. This interesting concept seems to be within the realm of possibility for small aberrant deposits but might be expected to obliterate the typical vitroclastic structure of welded tuffs. It might give rise to a streaky, banded flow rock which would have all external characteristics of a normal lava flow. The mechanism is conceivable in view of the existence of pyroclastic dikes and the possibility that such material could have been extruded *en masse* under optimum conditions.

This concept also is analogous to one held by Iddings as a mechanism for the formation of "welded pumice." Iddings (1909, p. 331) envisioned the extrusion as a lava flow of previously erupted fragmental materials that had fallen back into a vent and been reheated. He also explained flow breccia by this mechanism. Ross and Smith (in press) discuss this and other mechanisms of formation of the rocks Iddings called "welded pumice."

The controversy over the classical Armenian "tufflavas" and the Italian "piperno" has been discussed by Zavaritsky (1947, p. 15), who agrees with Paffenholz (1938) that the "tufflavas" are of pyroclastic origin, and with Dell'Erba (1892) and Zambonini (1919) that the Italian "piperno" was also a pyroclastic rock. Both "tufflavas" and "piperno" are regarded as lava flows by other authors (*e.g.*, Levinson-Lessing, Kalkovsky, and Abich, as cited by Zavaritsky).

Summary Statement

The evidence from published studies of welded tuffs overwhelmingly favors emplacement by avalanche or flowage of fragmental material and hot gas. None of the other proposed mechanisms has been convincingly demonstrated even for a minor deposit.

The development of the alternate theories for emplacement of welded tuffs however in part probably resulted from the reluctance of geologists (Grange, 1934; Kennedy, 1955; Hausen, 1954; Hentschel, 1955) to accept the "air-borne cloud" origin. This concept was an unfortunate misinterpretation of the mode of emplacement of nuées ardentes that undoubtedly was influenced by the devastating effects of the eruptions.

During the eruption of May 8, 1902, at Pelée it was the cloud that devastated St. Pierre and left behind ash and lapilli deposits up to a few feet thick (Anderson and Flett, 1903, p. 487). The avalanche from which the cloud was derived was probably confined to the valley of the Rivière Blanche. During the eruption of May 7, 1902, at Soufrière clouds caused the death and devastation of the settlement on both sides of St. Vincent and deposited a few inches of ash (Anderson and Flett, 1903, p. 448). The avalanches from which the clouds were derived were confined largely to the Wallibu and Rabaka valleys.

It has not been shown that a welded tuff

could ever form from the ash cloud. The size and temperature of Pelean-type clouds are probably inadequate. However, if we assume that clouds accompanied prehistoric ash flows, as they probably did, conceivably they were at times hotter and larger and may have carried more ash. If the glassy ash could have been *emplaced* at temperatures of the order of 900° C., it would most surely develop incipient or partial welding. Complete welding would depend on the thickness of the deposit. Such welded-tuff deposits, if they exist, might have interesting distributional and textural characteristics, but would be very subordinate in volume to the total products of the eruptions which produced them.

The supporters of the particulate-flow mechanism generally agreed that the driving force of the ash flow or avalanche is gravity, aided by expanding gases. Herein lies a problem that needs solution, namely, what is the true nature of the gas-emission process, and how does it vary in different ash flows? Is total vesiculation confined to the vent, conduit, or chamber, or does some of it take place in the flow, and, if so, how much?

If we assume that some vesiculation takes place during flowage, then gas is emitted from the flow in several ways: (1) release by vesiculation during flowage, which, if explosive, gives rise to the so-called "auto-explosive" property discussed by Perret (1935, p. 22–24) and Fenner (1923, p. 74); (2) escape of gas that was trapped by the sheer bulk of the ash and pumice, but which had been released by vesiculation before the flow formed; (3) rupture during flowage of vesicles that contain partly expanded gas; and (4) diffusion from the glass shards and pumice. Process (4) should take place as long as the glass remains oversaturated with respect to the temperature and pressure of its immediate environment but would not be "auto-explosive" in the strict sense. Vesiculation, especially in the vent, probably does not reduce the glass to a state of saturation or undersaturation, and the glass may continue to lose gas even after some cooling. If the amount given off by diffusion were only a few hundredths of a per cent by weight, the volume at the eruption temperature would be impressive and would provide an adequate lubricant.

This interplay of possible mechanisms could be important in determining the physical character of the final pyroclastic material. Some insight into the problem may be gained from statistical analysis of pumice and shard types and of sorting characteristics. For example, many ash-flow deposits contain quantities of pumice crowded with long fine, parallel to subparallel, tubular vesicles, and the shards may be seen to have formed by fragmentation of such pumice. Presumably the only place such pumice could form is in the conduit, and this suggests that most, if not all, vesiculation, together with much loss of gas, took place before the ash flow was formed. The lubricating gas of the ash flow, then, must be attributed to mechanisms other than auto-explosion in the strict sense. Some ash flows, however, show a paucity of long-tube pumice, and the shards are coarse irregular fragments, many with spherical vesicles. Other deposits are mixtures of types. Quantitative studies of shard and pumice types have not been made, and such studies may reveal important clues for the solution of ash-flow problems.

Recognition and Definition of Units

General Discussion

The recognition and definition of ash-flow units pose one of the most difficult problems of pyroclastic flows, especially in deformed rocks. Perhaps the principal difficulty relates to the fact that the basic unit, one ash flow, is, in many instances, locally indistinguishable in multiple-flow cooling units. The problem is made more difficult because cooling units, or single ash-flows, may be separable in two dimensions but sometimes not in three dimensions. This is, of course, a common geological dilemma, but it has special significance for welded tuffs, because the unique properties of these rocks permit new approaches to the solution of old, as well as new, problems.

The splitting of units, and even of the basic unit, is commonly a function of the scale of mapping or of the degree of understanding one wishes to achieve. Thus, for reconnaissance mapping, a lithologic unit may be a heterogeneous volcanic field, whereas for detailed mapping a unit may be a texturally distinctive zone of one welded ash flow. The following discussion may appear to be needlessly complicated for many small ash-flow sheets and for some larger ones; for the great caldera-forming eruptions, however, the discussion is probably an oversimplification.

Flow Units

The usual concept of a flow unit, stemming as it does from our knowledge of historic

nuée-ardente eruptions, is restricted because of the small size of these deposits and the fact that they are confined to valleys. The flow unit or avalanche is generally considered the product of one nuée ardente, and at Pelée, for example, successive nuées ardentes were confined to the avalanche valley because they were erupted from the dome through a breach or notch in the crater rim. The resulting deposit in the valley inevitably is a stack of thin layers or flows, separated from each other by block accumulations, fine ash, materials reworked by wind or water, or minor erosional disconformities. The nature of these partings may change from place to place between the same flow units, and locally a parting may not be readily detectable.

At Soufrière, because of the open crater, the distribution of the flows or avalanches was controlled largely by topography below the crater, and the materials poured down several valleys. Each of these valleys now contains a valley flow (presumably one from the May 7, 1902, eruption) which is a unit in itself, but all the valley flows are the product of a single eruption (except possibly minor material from the May 9 and 18 eruption).

Let us imagine a series of successive eruptions from a volcano in an environment similar to Soufrière. The first eruption may produce a group of independent valley flows. A second eruption produces an additional flow in each valley, but some of the flows may coalesce at the mouths of the valleys. As the surrounding ridges and valleys are buried by materials from successive eruptions and the terrain is leveled, the flows are no longer valley units but become successively larger, complexly lobate units. Finally they may become single sheet flow units continuous through 360° around the source. This model assumes approximately uniform radial distribution from the crater.

We may conceive of an eruption giving rise to a radially distributed single flow of petrographic homogeneity around small sources of the single-vent composite cone type. From larger source areas and from multiple vents such as those that culminate in large calderas and volcano-tectonic depressions, the pattern may progress with increase in dimensions and diversity of the source area to large and small overlapping and coalescing lobate sheet flows oriented radially in favorable sectors relative to their principal source vents. We should perhaps keep an open mind on the subject, but it seems incredible that a single lithologically homogeneous flow unit could be formed from multiple vents, or even one great fissure, to cover an area the size of, for example, the Toba field (2000 square miles). Such units probably represent something more complex than a single flow even though random sampling indicates homogeneity. On the other hand, because of the probability of the intermingling of materials from multiple vents, or continuous evolution of great volumes of material from one or more vents, our concept of a flow unit must embrace compound flows that could conceivably show nearly all possible variations seen in welded tuffs, except incompatible chemical compositions.

No description of a single flow unit in its remaining or exposed entirety has yet been published, and it is doubtful whether geologists will undertake such a task in the near future. It will thus be a long time before we penetrate the intricate maze of the large ash-flow sheets and understand their history and the history of their source areas.

Our immediate concern with the flow unit is in one- and two-dimensional vertical sections where it must be defined as the thinnest recognizable virtually nonsorted layer. Its thickness may range from inches to hundreds of feet, and it may be separated from other flows by partings of diverse characteristics (Smith, in press). The visual recognition of partings, and hence flow units, may be very difficult and locally impossible in some welded tuff sheets. Some work suggests that the magnetic properties of individual flows may be useful in distinguishing them (Hatherton, 1954a; 1954b).

The writer thinks that some of these flow units as seen in two dimensions are actually subflows that originated within the main flow itself, rather than at a common point of origin in the vicinity of the source. These subflows may result from the lateral overlapping of lobes on an irregular front or possibly by vertical overriding due to volume surges or deposition on an irregular surface. These are conjectural concepts, but the overlapping of lobes around topographic prominences can cause the appearance of two flows in vertical sections downslope from the prominence, whereas there is only one upslope from it. Whatever their origin, such units represent a physically distinct and tabular flowage mass, and we have little choice but to call them flows until detailed mapping proves otherwise.

Cooling Units

In deposits of welded ash flows the single ash flow is commonly subordinate to a unit that consists of more than one flow (Pls. 1–2). The flows, however, have been emplaced in close enough succession so that they cooled virtually simultaneously, or in continuum over an extended period of time.

These multiple-flow units may show zonal patterns of welding and crystallization indistinguishable from the pattern shown by single ash flows having approximately the same temperature and thickness characteristics. It is therefore commonly more practicable and less ambiguous to think of any zoned unit as a cooling unit: *simple* if it exhibits the zonation of a single flow unit, or *compound* if the zonal pattern deviates from that expectable in a single flow.

Because of the common difficulty of recognizing flow units within densely welded cooling units, probably many of the so-called single flows are really multiple-flow cooling units. An excellent example of this problem is shown by the "ignimbrite" sheet at Maraetai 5, New Zealand. Marshall (1935, p. 347) writes of this ". . . no discontinuity could be seen in it of such a nature as to suggest that there was more than one period of deposition." R. Bailey (Oral communication), accompanied by J. Healy, has re-examined this sheet and has observed discontinuities that show that at least three flow units are represented. Hatherton (1954b, p. 427) distinguished three "sheets" by visual examination, whereas his measurements of remanent magnetism suggested five probable "sheets." A similar situation may exist in the Motutere section described by Marshall (1935, p. 351), as is indicated by reversals within his measured density sequence (Fig. 5).

Fenner (1923, p. 29–31 and 1948, p. 885), Gilbert (1938, p. 1836), Williams (1942, p. 82), Westerveld (1942, p. 204), and others recognized this problem, but details are lacking, especially for the more densely welded and crystallized sheets where distinctions are more obscure. Many, and perhaps most, welded tuff sheets should be re-examined with this thought in mind.

Composite Sheets

It can be demonstrated that all degrees of welding can exist at the contacts between flow units within compound cooling units. The parting between two flow units may at one place occur in densely welded tuff, whereas elsewhere it may occur in nonwelded tuff. The specific character of these contacts in any given place is related to the time interval between flows, the emplacement temperatures and thicknesses of the flows, and the number of flows that constitute the cooling unit. Some combinations of these variables will cause the formation of units that will show zonations in vertical sections ranging from the patterns of simple cooling, through degrees of compound cooling, to separate cooling units—separate, that is, in so far as we are able to detect continuous cooling in a vertical section. In complex fields of long-continuing eruptions we may expect that single cooling units in one area will grade into separate cooling units elsewhere. These may be separated by erosional disconformities.

The aggregate of cooling units joined by welding or crystallization in some areas, but grading laterally into two or more separate cooling units, is called a composite sheet (Smith, in press). Because it has not been documented by mapping, although supported by inspection in a few localities and inferred in

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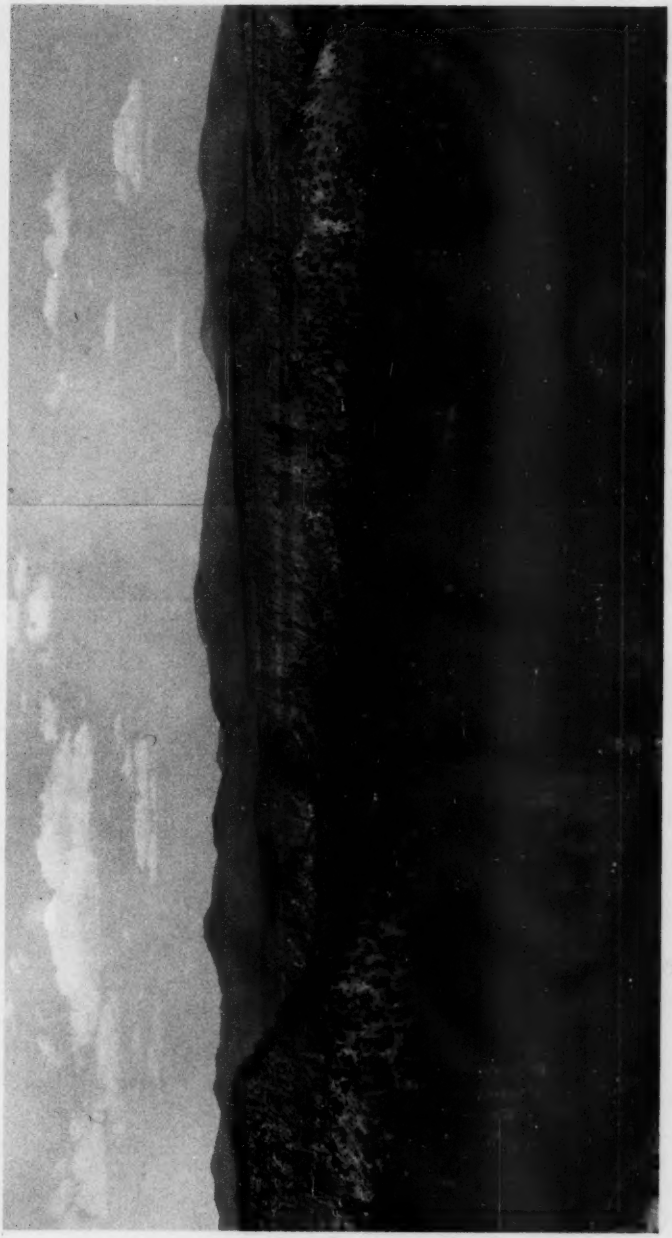
PLATE 1.—ASH-FLOW SHEETS ON THE PAJARITO PLATEAU NEAR LOS ALAMOS, NEW MEXICO

Two members, separated by an erosional disconformity and air-fall beds, make up the Bandelier rhyolite tuff in the photograph. The lower member is a nonwelded and vitric ash-flow unit that underlies the lower, tree-covered slopes of the mesas. The air-fall tuff and disconformity are visible at the far left, at the base of the overlying cliff-forming member. The upper member is a multiple-flow cooling unit whose partings are clearly emphasized by weathering. It is mildly welded and crystalline (vapor-phase zone) from the top of the mesas down to the middle of the lower dark band about midway up the exposed sections. The lower half of this band is a nonwelded, reddish, vitric zone. The degree of welding increases toward the mountains beyond which lies the source area. The partings between flow units become more obscure and locally invisible with increase in degree of welding. Photograph by R. A. Bailey.



ASH-FLOW SHEETS ON THE PAJARITO PLATEAU NEAR LOS ALAMOS, NEW MEXICO

FIGURE 1

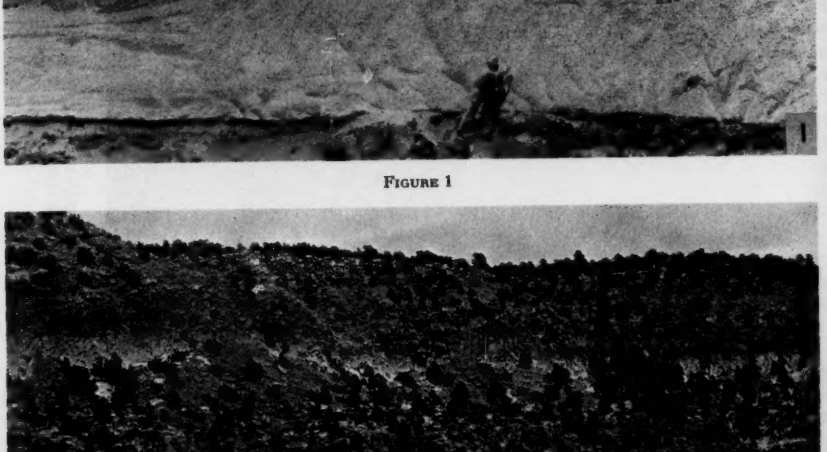


FIGURE 2

VARIABLE WELDING IN THE BANDELIER RHYOLITE TUFF

others, the unit is referred to as hypothetical until it can be accurately mapped. The existence of such a sheet seems inevitable, and it may eventually prove to be the most significant unit in terms of the eruptive and structural history of the source area, especially in deposits of large volume.

Additional problems of correlation may arise in ash-flow fields that are separated into sectors by topographic barriers. Here the composite sheet in one sector could conceivably have single-flow or separate cooling-unit counterparts in other sectors.

Ash-Flow Fields

An ash-flow field as here defined is the aggregate deposit of ash flows that can be related as a unit, or as a group of integral parts, to a specific source or source area. It may consist of any one or all of the other units either in single or multiple forms. The source is any simple volcanic vent such as a dome, crater, or small fissure, whereas source area refers to calderas, volcano-tectonic grabens, or other complex vent systems. Obviously such a flexible term cannot have meaningful genetic connotations, and it is used here only in a descriptive sense to provide some measure of order to an otherwise heterogeneous complex of volcanic rock deposits. One can envision subfields, superfields, multiple fields, and others, but such subdivisions call for data not available at present.

Ash-flow fields, in their entirety or in part, such as Toba (Westerveld, 1947), New Zealand (Marshall, 1935), Aso (Matumoto, 1943; Williams, 1941a), Chiricahua (Enlows, 1955), San Juan (Larsen and Cross, 1956; Ratte and Steven, 1959), Valles (Ross and Smith, in press), and many others, are stacks of cooling units or composite sheets with varying amounts of interbedded ash.

It is inferred that the greater the number of cooling units, and the greater the differences among them, the longer and more complex has been the eruptive history in the source areas.

Magnitude of the Deposits

Distance Traveled

The maximum distance an ash flow can travel from its source is not known. It may be of the order of 100 miles, although such a distance would be unusual in the light of existing data. Theoretically, temperature and gas content, hence the gas-emitting properties of an ash flow, appear to be the most significant variables influencing distance traveled. When actual distances traveled by specific ash flows are compared, however, gradient, topography, and especially volume appear to have exerted a greater control.

Figure 2 shows some maximum distances from known or inferred probable source areas. The most reliable figures for the greater distances are probably those given by Williams (1942, p. 79) for the "pumice flow" that can be traced down the Rogue River valley for 35 to 40 miles, and by Matumoto (1943, p. 4) who states that the farthest outcrop of the "Aso lava" is 100 km from the center of Aso caldera. The Nomlaki tuff of the northern Sacramento Valley, California (Anderson and Russell, 1939), may have traveled 50–65 miles or more from an unknown source on the eastern

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Plate 2.—VARIABLE WELDING IN THE BANDELIER RHYOLITE TUFF

Figure 1.—Three members of the Bandelier tuff along New Mexico Highway 4 near Totavi, New Mexico. Above a basaltic lava flow at the man's feet is a basal member of air-fall beds of lump pumice and ash of variable thickness. The nonwelded and vitric middle member (lower member in Pl. 1) underlies the tree-covered slope. The upper member is the multiple-flow cooling unit of Plate 1, crystalline and mildly welded from the base of the lower tier of columns to the top of the section. Denser welding near the top indicates compound cooling with a probable increase in emplacement temperature toward the top. Photograph by R. A. Bailey.

Figure 2.—North wall of Frijoles Canyon, Bandelier National Monument, New Mexico. Upper half of photograph equals upper member of Figure 1 of this plate. The darker layers indicate higher degrees of welding than the lighter. The lower light-colored cliffs are formed in nonwelded materials and are pock-marked with wind-eroded holes. Prehistoric Pueblo Indian ruins at their base are a feature of nonwelded units of the Bandelier ash flows in the Jemez Mountains. Photograph by George A. Grant, courtesy of the National Park Service.

side of the valley to the foothills of the western side. On the western side of the valley the tuff sheet has an outcrop area of more than 40 miles in a north-south direction.

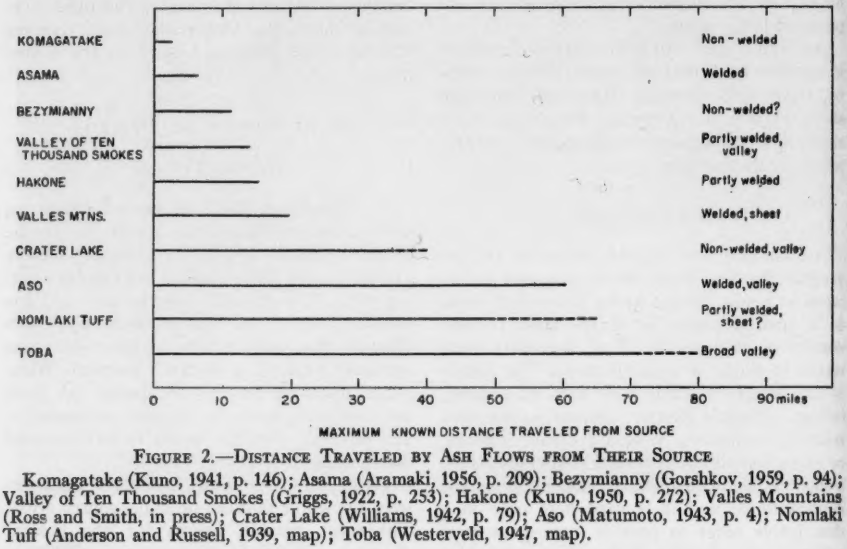

FIGURE 2.—DISTANCE TRAVELED BY ASH FLOWS FROM THEIR SOURCE

Komagatake (Kuno, 1941, p. 146); Asama (Aramaki, 1956, p. 209); Bezymianny (Gorshkov, 1959, p. 94); Valley of Ten Thousand Smokes (Griggs, 1922, p. 253); Hakone (Kuno, 1950, p. 272); Valles Mountains (Ross and Smith, in press); Crater Lake (Williams, 1942, p. 79); Aso (Matumoto, 1943, p. 4); Nomlaki Tuff (Anderson and Russell, 1939, map); Toba (Westerveld, 1947, map).

The distance of 80 miles shown for the Toba tuff represents the most distant outcrop of tuff, shown on Westerveld's (1947) map, measured from the northwest rim of the Toba depression down the L. Biang and S. Wampoa valleys. This is tentative, however, because the most distant outcrop may represent fall deposit rather than flow.

Many deposits of ash flows are continuous for more than 50 miles, but in the absence of a known source area all such distances probably represent only a partial dimension of a large sheet. Mackin (1960, p. 95) and Cook (1958, p. 1548) have mentioned single welded-tuff units that extend for more than 100 miles in southwestern Utah and southeastern Nevada. If this represents distance from a source it is a prodigious figure, but if it represents diameter or long dimension of a single sheet it is large but not excessive when we consider, for example, the Toba tuff which is a continuous map unit for more than 150 miles. Admittedly this may not be a fair comparison, because the individual units of the Toba tuff have not been distinguished. However, the comparison does emphasize the large size of some welded-tuff deposits.

Area

The area covered by ash-flow fields is, of course, a function of the volume of the deposit and its thickness; thus the topography surrounding a source area influences the thickness of the tuff, its area, and its symmetry. Figures generally quoted for area and volume are for *pyroclastic fields* which may or may not (ordinarily not) consist of one ash flow. Usually the eruptive history of any field is so complex that detailed mapping has not been done and the number and kind of units are not known.

It is therefore easy to overlook the fact that when the magnitudes of such ash-flow fields as Crater Lake and Aso are compared, they are not geologically equivalent. The Crater Lake deposits (Williams, 1942) appear to represent one catastrophic eruptive period and thus one unit, which, notwithstanding the fact that it probably consists of many ash flows, appears to have been a unit of cooling. The Aso deposits, on the other hand, comprise at least three if not five or six major cooling units and thus represent several periods of eruption separated by at least one marked disconformity (Williams, 1941a; R. A. Bailey, oral communication). Conditions similar to or more complex than those at Aso are common in a great many ash-flow fields surrounding calderas. They show the complexity of formation of these fields and associated calderas.

The known areas and volumes of specific ash-flow fields have been partially tabulated by several writers (Westerveld, 1942; 1952; Weyl, 1954; Rittmann, 1958, p. 525). Some of these data for volumes are incorporated in Figure 3. The most interesting and geologically important ash-flow fields cover several hundreds to thousands of square miles. Some of them contain zones of densely welded tuff throughout their entire known area. Few, if any, are known in great detail, and many years of work will be necessary to determine their volume, source, number of eruptive cycles, and tectonic relationships.

Fields most often cited as the world's largest are those of the Lake Toba region, Sumatra, and of North Island, New Zealand. Here again comparison is difficult because of basic differences in the characteristics of each field. Although these two are probably better known than others, they are as yet not really well studied. Toba as deduced from the writings of van Bemmelen (1949) and Westerveld (1947; 1952) appears to be enormous but not greatly complex when compared to New Zealand. The ash-flow sheets seem to have originated within a relatively short period of time from the area now occuped by the giant Toba volcano-tectonic depression. The New Zealand field, on the other hand, seems to be made up of ash-flow sheets from numerous sites, erupted over a rather long span of time. As a consequence the subsidence history of the Taupo-Rotorua graben in New Zealand seems far more complex than that of the Toba depression. Other large fields are so little known that we can only record their existence.

YELLOWSTONE ASH-FLOW FIELD: Boyd (1957, Ph.D. Thesis, Harvard Univ.) has recently estimated that Yellowstone National Park contains 600 square miles of welded tuff exposed within the park; added to the probable western extension of the field this gives a figure of 2000 square miles. Addition of an equal area presumed to underlie the plateau rhyolite flows and the Snake basalts gives a total of 4000 square miles. Within the park the thickness ranges from 500 to 1000 feet. With 500 feet as an average original thickness and 4000 square miles as probable areal extent, Boyd calculated that the volume of welded tuff would be on the order of 400 cubic miles. He suggests that it was all emplaced in a single rapid sequence of eruptions.

ELKHORN MOUNTAINS ASH-FLOW FIELD: One of the largest ash-flow fields known is part of the Elkhorn Mountains, Montana, volcanic field of Late Cretaceous age which surrounds and was intruded by the Boulder batholith. This field appears to consist of several thick and extensive welded-tuff sheets which occupy a minimum area of 3000 square miles and have a minimum volume of 500 cubic miles, but which may have an area as great as 10,000 square miles and a maximum volume of 1000 cubic miles (Klepper and Smedes, 1959, p. 1631; and oral communication).

SAN JUAN MOUNTAINS ASH-FLOW FIELD: One of the most spectacular and complex ash-flow fields in the world is associated with the San Juan Mountains volcanic pile in Colorado and northern New Mexico. Of specific interest are three of the upper members of the Potosi series—the Treasure Mountain, the Alboroto, and the Piedra. Larsen and Cross estimate total volume for these as 2350 cubic miles. From the work of geologists of the U. S. Geological Survey, including Burbank, Larsen, Luedke, Olson, Ratte, Ross, Steven, and others, and evaluation of field data over a period of 15 years by R. A. Bailey and the writer, it can be said that most, if not all, of the three members are silicic pyroclastic rocks, in large part welded tuff. The area covers approximately 12,000 square miles. At present only one source area, the Creede caldera (Steven and Ratte, 1959; Ratte and Steven, 1959), is definitely known. Two other calderas, the Silverton and Lake City (Burbank, 1941; Burbank and Luedke, oral communication, 1959), are seemingly related to the older rock series among which ash flows are also known. The field should be classed as a multiple-source field; other source areas remain to be found and related to specific salients of the field.

SUDBURY BASIN. On the basis of recent work in the Sudbury basin, Ontario, Canada, Thompson (1957, p. 43) and Williams (1957, p. 69, 70) have postulated that the Sudbury "lopolith" is really a volcano-tectonic depression surrounded by a ring of eruptive centers and that the subsided basin contains no less than 300 cubic miles of "glowing-avalanche" deposits. Deposits about a mile thick are believed to occupy about 300 square miles within the basin.

BASIN AND RANGE ASH-FLOW FIELDS: Another tremendous, and probably very complex, ash-flow field occupies an area of more than 10,000 square miles in southwestern New Mexico. It comprises large parts of the Datil, Mogollon, Black, Tularosa, and other mountains and probably has a volume equal to or greater than

the San Juan field. This field is a small part of the vast volcanic system of Tertiary age of the Basin and Range province of western United States and Mexico.

Welded tuffs in the Basin and Range province have long been known, although Ross (1953; 1955, p. 430–432) seems to have been the first to make a clear statement in the literature. Recently Williams (1957, p. 64) has stated that not less than 50,000 square miles of southern Nevada and southwestern Utah was buried to a maximum depth of about 8000 feet by "fissure-erupted glowing avalanches" during late Tertiary time. Mackin (1960, p. 83) has estimated the volume of silicic volcanic rocks in the Great Basin to be about 50,000 cubic miles, and this is probably conservative.

Thickness

Many figures have been published for thicknesses of "welded-tuff sheets," "ignimbrite sheets," "glowing-avalanche deposits," "tuff flows," and related rocks. What these thicknesses mean in terms of the units defined in this paper is problematical. For many general problems the lack of detail is not serious, but it is critical to many genetic problems if units of equal rank must be compared. Perhaps most published figures are for aggregate tuff deposit or deposits, and figures for "a welded-tuff sheet" or "an avalanche deposit," or "one ignimbrite" may commonly represent multiple-flow cooling units.

Marshall (1935, p. 328) found that the New Zealand "ignimbrites" range from 60 to 500 feet in thickness but stated that many are less than 100 feet thick. Some are probably single-flow units, but we now know that others are multiple-flow cooling units. Jenks and Goldich (1956, p. 162) state that locally the "white-tuff flows" are 160 feet thick, whereas three successive tuff flows composed of salmon sillar, with very little interbedded material, aggregate 500 feet.

Enlows (1955, p. 1221) gives figures ranging from 2½ to 880 feet for the seven or eight "members" of the Rhyolite Canyon formation, which he considers welded tuff. He infers (p. 1219) that each layer or "member" is the product of one "nuée ardente" or "hot-ash" eruption. Member No. 6 contains 700 feet of densely welded tuff. Mansfield and Ross (1935, p. 312) found that their Idaho welded tuff ranges in thickness from 20 to 50 or more feet.

Individual pyroclastic flows evidently range from a few inches to hundreds of feet in thickness. Their individual character is probably determined primarily by their volume and by topography, although we might expect that, if volume and topography were the same, the hotter, finer-grained flows would be thinner than the colder, coarser-grained ones, although this difference might be influenced by gas content. Other variables might be considered, but at any rate we must expect that small-volume, short-burst eruptions from "Pelean" domes will result in thinner flows than will larger-volume, continuous-evolution-type eruptions from craters or fissures. Nevertheless the latter type may also result in thin flow units. Most multiple-flow cooling units probably have a similar range in thickness but have a higher average thickness than do single flow units. Exceptionally thick cooling units are known.

The thickest welded-tuff sheet that the writer would interpret as a cooling unit is the Superior dacite in the Globe-Miami-Superior area, Arizona. This welded-tuff sheet was kindly shown to him in 1956 by N. Peterson and in 1957 by D. Peterson, both of whom have studied parts of it in some detail. The Superior dacite was originally described by Ransome (1903) as a lava flow more than 1000 feet thick, and again by N. Peterson, Gilbert, and Quick (1951), who discuss the possibility of its pyroclastic origin. D. Peterson (1959, p. 1740) presents the evidence for a pyroclastic origin and suggests that the sheet, which covers at least 400 square miles and is locally as much as 1500 feet thick, "was deposited by a single huge pyroclastic eruption." The precollapsed thickness of this sheet was probably well over 2000 feet. Publication of the details about this most interesting welded-tuff sheet will add greatly to our understanding of the problems.

Thicker cooling units may exist in the San Juan Mountains of Colorado. Those currently being studied by Steven and Ratte (1959, p. 1788) in and around the Creede caldera are of special interest. Ratte and Steven (1959, p. 1785) report two major stages of ash-flow deposition: (1) a precaldera rhyolitic stage forming deposits more than 3500 feet in thickness and ranging from densely welded tuff "resembling fluidal rhyolite in the lower part to essentially nonwelded pumice breccia at the top", and (2) a caldera-subsidence stage presumably culminating in more than 4000 feet of quartz latitic welded-ash flows that are now exposed in the caldera core. Other lesser ash-flow, pyroclastic, and lava-flow units occur within the volcanic edifice.

Stacks of cooling units in the Basin and

Range province in Nevada reach a known maximum of 8000 feet (Williams, 1957, p. 64). Such thicknesses probably represent overlapping fields which, in aggregate, are even thicker.

Volcanic Vents and Source Areas

The most critical problem of ash flows, especially the large-volume sheet deposits, concerns the location and characteristics of the source areas.

Domes

With few exceptions, historic nuées ardentes have been erupted from volcanic domes and have produced deposits of small volume and restricted distribution. None of the deposits from domes is known to have formed welded tuffs although incipient welding in some of these deposits may have been overlooked. Even though future studies show that some eruptions from domes resulted in the formation of welded tuffs this should in no way obscure the fact that such deposits will be rare, relative to welded tuffs of other origins.

The best-known examples of nuée-ardente eruptions from domes are the 1902 and 1929–1932 eruptions of Pelée, Martinique, and the periodic eruptions from Merapi, Central Java. More recent eruptions were Mt. Lamington, New Guinea (Taylor, 1954; 1956), Hibok-Hibok, Philippines (Macdonald and Alcaraz, 1956), and Bezymianny, Kamchatka (Gorshkov, 1959).

Craters

The classical example of a nuée-ardente eruption from an open crater is the 1902 eruption of Soufrière, St. Vincent. The resulting ash-flow deposits were seemingly unconsolidated. Incipient welding was probably precluded, not because of too low a temperature, but because of the high-crystal, low-glass content and inadequate thickness.

Welded tuffs have been produced by historic eruptions from open craters, although the few examples are of types far removed from rocks normally called welded tuffs. Aramaki (1956) has described the "Agatsuma nuée ardente" as one of the products of the 1783 activity of Asama volcano. One lobe of this flow reached 8.5 km from the crater, its total area is estimated at 18.5 km, and its volume (assuming an average thickness of 5 m) is approximately 0.1 km^3. The thickness ranges from more than 10 m in the central part to less than 4 m at the edges.

Fissures

Small central vents can undoubtedly produce welded tuffs, but most recent writers believe that the large-volume ash-flow sheets (tens of cubic miles) have been erupted from fissures (Williams, 1957, p. 64; Westerveld, 1952, p. 564–568; van Bemmelen, 1949, p. 202; Ross and Smith, in press). Some writers have postulated, in the absence of positive evidence for fissures, that the fissure vents for these large-volume deposits are buried beneath the ash-flow sheets (Cotton, 1952, p. 211; Weyl, 1954, p. 25).

No evidence unequivocally supporting such a concept has been published. Several writers have found fissure dikes which they thought were feeders, but none has proved it. As indicated above, the writer is speaking of deposits with a volume in excess of several cubic miles. This would exclude the deposits of the Valley of Ten Thousand Smokes deposit which Williams, Curtis, and Juhle (1956, p. 129) now believe to have been erupted from fissures at the head of the Valley as well as from the vent now occupied by Novarupta. Fenner (1923) held much the same view, although he also suggested fissures on the Valley floor as a supplementing source. As pointed out by MacGregor (1952) Fenner has often been misquoted on this point.

The New Zealand ignimbrites are often cited as fissure-erupted materials, and the "positive" evidence seems to center around an "ignimbrite dike" which outcrops at Ongarue, and which is pictured by Cotton (1952, p. 210, Fig. 102). The writer has been skeptical of the "feeder dike" interpretation for many years and was interested to learn from Roy A. Bailey (oral communication, 1957) who has examined the outcrop in the field, that there is indeed some doubt as to the intrusive origin of the "dike". It shows a horizontal planar eutaxitic foliation just like that seen in the surrounding ignimbrite rather than the vertical lineation expected in a dike. Bailey suggests that the horizontal jointing may be related to an abnormal cooling surface such as might be found if an ash-flow filled a narrow gorge. Detailed study is needed. Aside from this one doubtful locality it would appear that no occurrence is known where pyroclastic dikes actually project into ash-flow sheets where it can be shown or rea-

sonably inferred that the dikes were actually feeders for the large-volume deposits.

The evidence for fissure feeders seems established for at least part of the Valley of Ten Thousand Smokes deposit (Williams, 1954) and for an extremely interesting occurrence in the John Day formation of Oligocene age in Oregon, kindly shown the writer by A. C. Waters. In the latter area a dike several miles long appears to change from fluidal rhyolite in its deepest exposed part to fluidal-fragmental welded tuff near the surface and merge with a densely welded tuff sheet on the surface.

Other areas where fissure feeders are suspected include the Pilomasin basin, Sumatra (van Bemmelen, 1949, p. 681), Macdoel quadrangle, California (Williams, 1949, p. 32), Sudbury, Ontario (Williams, 1957, p. 64), Independence Pass, Colorado (Burbank and Goddard, 1935), and Black Range, New Mexico (Kuellmer, 1954).

The negative evidence favoring fissure vents for the large-volume ash flow sheets is inconclusive but nevertheless more suggestive:

(1) Specific central vents as source vents seem as rare as fissures

(2) Apparent absence of volcanic cones in association, in many areas, with ash-flow sheets

(3) Linear arrangement, in some areas, of post-ash flow vents and intrusive rocks

(4) Post-ash flow collapse along arcuate fractures or other linear zones of weakness producing calderas and volcano-tectonic depressions

It is significant that nearly all ash-flow fields with a volume of more than a few cubic miles and that have been related unequivocally to source areas are associated with calderas or other depressions of subsidence. The specific vents of emission of the ash flows no longer exist at the surface; they were destroyed by the collapse after or during eruption. Those vents that were not destroyed or that may be present in the collapsed blocks are probably modified by structural readjustments, post-collapse volcanism, or intrusion, and are no longer recognizable. In deeply dissected volcanic and volcanic-plutonic structures conduits may have been obliterated by ring dikes, stocks, or other intrusive bodies.

Calderas and Volcano-Tectonic Depressions

Subsidence related to igneous activity has long been recognized, but van Bemmelen (1930) is probably to be credited with the theory of catastrophic engulfment of volcanic edifices after the eruption and draining of magma chambers of large volumes of pyroclastic material. Williams (1941a) and others have reiterated and amplified this view, calling attention to the fact that many of the pyroclastic deposits surrounding such collapse structures are of "nuée-ardente" origin, and that many of these deposits contain welded tuffs.

Specifically, when we find welded tuffs or other ash-flow deposits, where cooling units or composite sheets can be shown to have volumes of a few to many cubic miles, we may expect to find subsidence structures in the source areas. The existing positive evidence for young (Pleistocene and Recent) deposits supports this view. The characteristics of these structures may vary, but probably caldera-type structures predominate.

Figure 3 shows volumes for a selection of pyroclastic flow deposits. The logarithmic method of plotting gives undue weight to differences in the smaller deposits and suppresses more significant differences in the larger deposits, but it also allows a grouping of deposits on a small-scale plot which has geologic meaning.

This breakdown by volume is not an attempt at classification in the sense of Aramaki's (1957, p. 21–31) important three-fold classification of pyroclastic flows, but rather an attempt to generalize the interrelationship of volume of deposits with types of source areas. It is especially designed to show that where source areas are known, and well studied, for deposits of large volume these areas are sites of volcanic or volcano-tectonic subsidence, a fact which should be of critical importance, especially to students of older rocks.

The figures for volumes shown in Figure 3 are the best available. Unfortunately it is problematical if many can be compared as equivalent entities. The figures represent units of all rank between single-flow units and complex pyroclastic fields and are not converted to equivalent porosities. However, if we take cooling units or composite sheets as here defined as the units to be compared, it is doubtful if subdivision of the larger fields or more accurate estimates of volumes of the deposits will change the figures more than one order of magnitude.

The data suggest that the maximum volume for deposits erupted from volcanic domes, and hence Pelean-type deposits, strictly speaking, is in the fourth order of Figure 3, or somewhere between 1 and 10 km^3. Most of these deposits have a volume less than 1 km^3. None is known to contain welded tuffs. The term "Pelean" should probably be restricted to deposits of

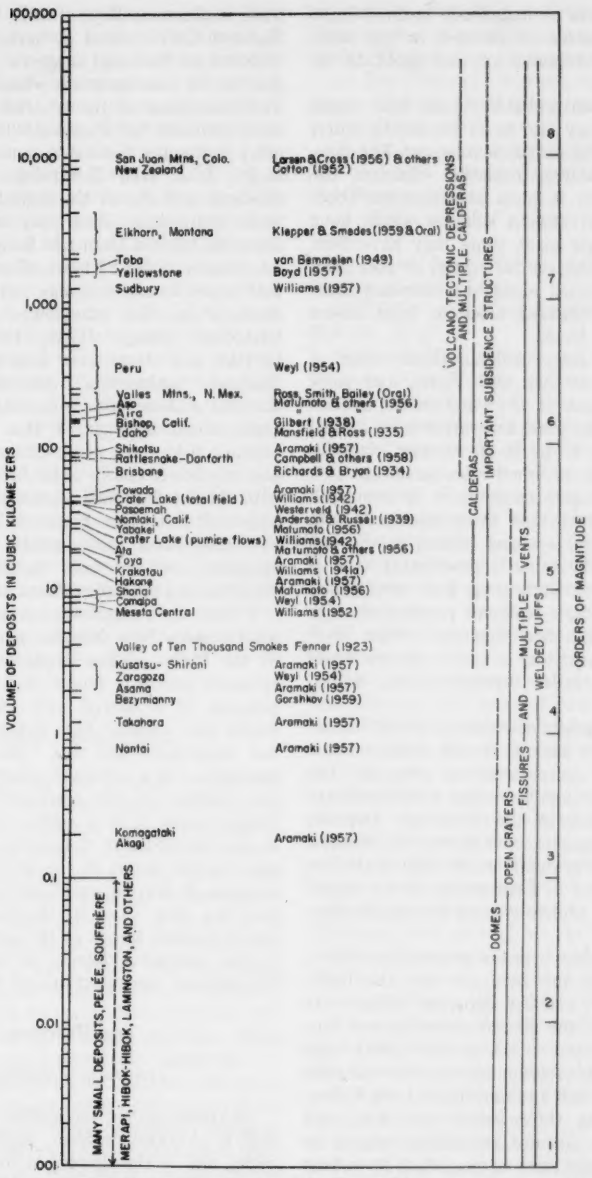

FIGURE 3.—VOLUMES OF PYROCLASTIC-FLOW FIELDS

The volumes, in km³, are plotted on five-cycle logarithmic paper and span seven orders of magnitude from 0.001 to 10,000 km³. The data show a correlation with vents and source areas.

these low orders of magnitude erupted from domes, irrespective of whether or not their emplacement mechanics are also applicable to other deposits.

The maximum probable volume from single open craters may also be in the fourth order, although the fifth cannot be ruled out. The overlapping of caldera formation obscures the nature of vents in most examples described. Deposits found around calderas which have formed in single giant cones may have been erupted from the central crater, or they may have been erupted along the boundary fractures of the subsided block or from fissure systems in the block.

All known large calderas (sixth order of Fig. 3) such as Aso and Valles, and some smaller ones such as Toya and others, are not related to single cones but rather were formed by subsidence of parts of volcanic plateaus, groups of cones, or broad volcanic shields. It is difficult to imagine single vents or even multiple vents, other than those associated with fracture (fissure) systems related to the collapse itself. The same is presumably true of many subsidence structures, both smaller and larger than those calderas associated with sixth order ash-flow deposits, which were guided by pre-existing or active tectonic trend lines and are called volcano-tectonic depressions.

Ash-flow deposits of orders 5, 6, and 7 seemingly either are associated with collapse structures or their source areas are unknown. The writer has been unable to find a well-authenticated exception in the literature. Deposits from large composite cones such as Mt. Mazama (Crater Lake) evidently reach their maximum volume in order 5. They surely do not exceed 100 km^3, and probably most are smaller than 50 km^3.

Of the ash-flow deposits grouped in order 6, the sources of the Brisbane tuff, the Idaho "welded tuff", and the Peruvian "sillars" are unknown, and the volume estimates are very tentative. Gilbert (1938, p. 1859–1860) suggested a possible relation between the extrusion of the Bishop tuff and faulting in Long Valley. The remaining three fields—Aso, Aira, and Valles (order 6)—are genetically related to structures of the same names which have been called the world's largest known calderas. Each exceeds 12 miles in diameter.

The ash-flow fields of order 7 have been mentioned briefly in the section on Area. Thompson (1957) and Williams (1957) consider the Sudbury basin to be a volcano-tectonic depression; Boyd (1957, Ph.D. Thesis, Harvard Univ.) noted a "tectonic basin" in Yellowstone Park and suggested a caldera origin, but he was uncertain whether subsidence was consequent on the eruption of 400 cubic miles of welded tuff or associated with 200 cubic miles of rhyolite flows that now nearly fill the basin. Toba (van Bemmelen, 1939) is the classical, and one of the largest, volcano-tectonic depressions. It is only one of several along the Barisan Mountain Range in Sumatra. The source of the Elkhorn, Montana, ash-flow field is not known and may have been totally obscured by the intrusion of the Boulder batholith; Grange (1934; 1937), Williams (1941a), and others have considered the New Zealand "ignimbrites" related to Taupo-Rotorua graben which is considered to be volcano-tectonic in origin. Within the San Juan volcanic field three giant calderas are known, but only one is known to be directly associated with the ash-flow field discussed on an earlier page and plotted on Figure 3. The volume of the field seems incompatible with a single structure, and the field should probably be classified as a multiple-caldera field.

Within the tremendous ash-flow complex of southwestern New Mexico, mentioned earlier, lie the Plains of San Augustin, a grabenlike structure probably larger than the Toba depression. It is floored with Pleistocene sediments and volcanic ash. Griggs (1954, p. 3) has suggested that this "prominent closed depression" is a collapsed structure related to the eruption of the associated welded tuff. Griggs reports beds of welded tuff downfaulted to low elevations in the interior of the plains and rhyolite domes in the interior and at the margins. It may be significant that this structure lies very close to the junction of two major tectonic trends at the southeast margin of the Colorado Plateau; its long dimension lies athwart the directions of these trends.

The Deposits

Introductory Statement

As stated earlier, recognition of welded tuff still is a major problem, especially in older rocks, and in the absence of clear-cut geologic relations it is not always possible to distinguish welded tuffs from lava flows. Many writers have considered criteria for recognition of welded tuffs, notably Marshall (1935), Gilbert (1938), Weyl (1954), Enlows (1955), Ross and Smith (in press). No single criterion is in itself

infallible. The ultimate criteria are those that establish that the rock is or was composed of vitric, at least in part, pyroclastic materials, chiefly glass shards and pumiceous fragments, and that these have been deformed and welded while they were still hot. The writer has found that the most pertinent observations are usually made at or near the base of the unit in question. Here the transitions from nonwelded to densely welded and from glassy to crystalline rock are within a few feet, and the physical changes in the fragmented materials may be easily traced, in contrast to the more gradual changes in the upper part of the sheet. Even after some metamorphism, relict vitroclastic structure may still be present in what was formerly the densely welded glass zone. The base of a welded-tuff sheet does not show the basal breccia zone normally found in silicic lava flows.

Even more difficult than the distinction between some welded tuffs and lava flows is the distinction between nonwelded ash flows and tuffs of other origins. Voluminous pyroclastic fall materials, near source areas, as well as avalanches of loose accumulations of pyroclastic materials may be indistinguishable from true pyroclastic flows, although the origin of most such deposits is decipherable if they can be traced laterally.

Sorting Characteristics

Ash-flow deposits are generally nonsorted, as can be seen in any outcrops of nonwelded or poorly welded materials; fragments range in size from blocks to dust. The chief differences among deposits seem to be in the proportions of coarse to fine materials and in the relative proportions of crystals and glass. Before welding or crystallization, most ash-flow deposits are composed of (1) glass shards and glass dust, (2) pumice or scoria fragments and occasionally bombs, and (3) crystals; glass generally predominates over crystalline materials. Most contain some accidental lithic fragments.

The existing data do not indicate that major sorting is ever an important phenomenon of the flowage mechanism itself. Probably, however, heavy accidental inclusions, other than those locally derived, will decrease in size and amount from source to distal end of a given flow. The amount of fine-grained materials should probably also increase away from the source, especially in flows of low gas content.

Figure 4 shows cumulative grain-size curves for localities where size analyses are available. Although few in number, they are remarkably similar. The following localities are represented:

(1) The "Shirasu" deposits, South Kyushu, Japan (Taneda, 1954; 1957; Taneda, Miyachi, and Nishihara, 1957)

(2) Pumice deposits at Komoro on southwest foot of Asama volcano, Japan (Tsuya, Murai, and Hosoya, 1958)

(3) "Older pumice", Crater Lake (Moore, 1934)

(4) "Sand flow", Valley of Ten Thousand Smokes (Fenner, 1923)

(5) "Sillar" near Arequipa, Peru (Jenks and Goldich, 1956)

(6) Nonwelded portion of "ignimbrite", Arapuni, New Zealand (Marshall, 1935)

The "Shirasu" and the Komoro deposits are by far the best studied for their sorting. The "Shirasu" curves, Sh-35 and Sh-60, represent the highest and lowest median (size at 50 per cent) values of nine samples considered by Taneda and others to be representative of the bulk deposit. The Komoro curves, K-5 and K-8, also represent the highest and lowest median values of nine samples considered to be representative of the bulk deposit, and curve K-10 represents the fine-grained upper layer of the deposit.

The Crater Lake curves, C-34 and C-53, are the coarsest and finest of four samples but cannot be too closely compared with the others, because they do not include material larger than 64 mm in diameter. As a result their median particle sizes are larger, and their total contents of ash-size material are somewhat smaller than shown here.

The "sand-flow" and the "sillar" curves are not adequate representation of their respective deposits, but neither can they be much in error. The same may be said for the Arapuni, New Zealand, curves, which are both from the same "ignimbrite sheet", one from above and the other from below the welded zone.

The significant thing shown by these curves is the preponderance of ash and fine ash-size materials. In all the samples plotted 70 per cent of the material is smaller than 4 mm. The median sizes lie between 0.08 and 0.8 mm, or close to one order of magnitude. As mentioned above, at least some of the Crater Lake material will be generally coarser if the sizes above 64 mm were included.

The data, few as they are, are a surprisingly good representation of ash flows. The "Shirasu" is a deposit of large volume associated with the

"hai-ishi" (welded tuff) which surrounds the giant Aira caldera. The "older pumice" is a nonwelded deposit associated with the formation of Crater Lake caldera in a large composite andesitic volcano; the Komoro deposit is a small-volume nonwelded "pumice flow" presumably from the summit crater of Asama volcano, an andesite cone. The "sand flow" of the Valley of Ten Thousand Smokes, a small-volume deposit erupted from fissures and perhaps a crater, is at least partly welded and is partly crystalline. The Peruvian "sillars" are from a large-volume deposit of unknown origin and consist of nonwelded, mildly welded, and crystalline facies. The Arapuni samples represent a finer-grained type of the fissure-erupted "ignimbrites" which form the vast tuff deposits of the North Island of New Zealand.

The writer offers with considerable confidence the opinion that *most* welded tuffs have been derived from ash flows having sorting characteristics within the range of the cumulative curves of Figure 4. Some, nevertheless, will contain a higher proportion of coarse blocks of pumice or scoria, and a few such as the welded tuffs from Idaho, studied by Mansfield and Ross (1935), and the Walcott tuff (Stearns and Isotoff, 1956) are probably finer-grained.

FIGURE 4.—CUMULATIVE CURVES SHOWING SORTING IN ASH FLOWS

SH curves: "Shirasu" deposits, South Kyushu, Japan (Taneda, 1954; 1957; Taneda, Miyachi, and Nishihara, 1957)
A curves: Asama volcano, Komoro deposits, Japan (Tsuya, Murai, and Hosoya, 1958)
CL curves: "Older pumice," Crater Lake, Oregon (Moore, 1934)
VTTS curve: "Sand flow," Valley of Ten Thousand Smokes, Alaska (Fenner, 1923)
P curve: "Sillar" near Arequipa, Peru (Jenks and Goldich, 1956)
N.Z.A. curve: Nonwelded portion of "ignimbrite," Arapuni, New Zealand (Marshall, 1935)

Seemingly there is no obvious correlation between sorting and volume of deposit, composition, type of vent, or source area. The data, therefore, do not supply criteria for separating ash flows erupted from craters from fissure eruptions. Statistical studies of a large number of deposits might show that there is a tendency for rhyolitic or fissure-erupted material to be finer-grained than others, but even this seems doubtful. The principal differences among the curves shown in Figure 4 seem to be in the variation in amount of coarse materials (pumice fragments above 4 mm) both within a single deposit and in different deposits.

The data suggest that even within the framework of the cumulative curves (Fig. 4) there is considerable room for variation, and detailed studies from many deposits should yield use-

ful data. If, for example, a variable vertical component plays a significant role in these eruptions, then we have a mechanism that can influence sorting features *before the ash flow is formed*. Crystal enrichment or, rather, fine glass shard and dust impoverishment might take place in the eruptive column giving rise to ash flows that are not strictly representative of the magma from which they were formed. Comparative studies of bulk crystal–glass ratios in the flows, and of crystal–glass ratios in the pumice fragments should be made wherever possible. Studies contrasting nonwelded ash flows with the nonwelded facies of densely welded gas-rich ash flows might supply information on the role of gases in reducing attrition and sorting during emplacement, and might also further our understanding of the pyroclastic-flow mechanism.

A few writers have considered attrition during flowage to be responsible for the high proportion of fine ash in these deposits. Whether or not this is true remains to be demonstrated. It does not seem possible that such a mass of fragmented material could flow for miles without considerable attrition (especially near the distal end), yet if attrition does take place there should be a marked increase in amount of ash and fine ash from source to distal end of the flow. There are, as yet, no really diagnostic data bearing on this facet of the problem. Although Moore's (1934, p. 365) data suggest that there is an increase in material smaller than 4 mm away from the source, in the Crater Lake pumice flows this may be due to sorting. Williams (1942, p. 82) states that the size of pumice lumps bears no relation to distance from the source although they are generally more rounded at the distal ends. Krumbein and Tisdel (1940, p. 304) have shown that the sorting characteristics of the Crater Lake "pumice flows" conform to Rosin's law of crushing, a fact which may have some bearing on our interpretation of the pyroclastic-flow mechanism. It probably remains to be determined whether explosive disintegration and attrition produce the same sorting characteristics. The sorting characteristics of lahar (volcanic mud flow) deposits are very similar to those of ash flows, but perhaps of generally coarser grain.

Welding

The sticking together, cohesion, or welding of fragments of tuffaceous rocks was recognized by Fritsch and Reiss (1868, p. 414), Zirkel (1876, p. 267), and Dell'Erba (1892) long before Iddings (1899) introduced the term "welded pumice." The term "welded tuff", commonly incorrectly credited to Iddings, probably evolved from Iddings' use of "welded pumice." It was first formally introduced by Mansfield and Ross (1935), although Ross (oral communication) says that "welded tuff" was in use in the U. S. Geological Survey long before 1935. His first recollection of using the term was in 1919 in connection with rocks in the eastern San Juan Mountains near Crestone, Colorado. Their origin was then a puzzle, but we now know from re-examination that these rocks are densely welded tuffs of ash-flow origin.

The welding process as it applies to ash-flow materials promotes the union or cohesion of glassy fragments in a viscous state. The degree of welding may range from incipient stages marked by the sticking together or cohesion of glassy fragments at their points of contact to complete welding of the surfaces of glassy fragments accompanied by their deformation and by the elimination of pore space, and perhaps ultimate homogenization of the glass.

Phenocrysts and rock fragments are rarely affected by the welding process nor do they have much effect on it, except in rare instances where they are so abundant as to inhibit consolidation of the glass fragments.

Many variables affect the degree of welding; the more important ones are (1) temperature, (2) amount and composition of volatiles, (3) composition of ash, (4) lithostatic load, (5) rate of cooling, and (6) rate of crystallization. The first three control the viscosity of the glass and for purposes of discussion may be reduced to the one variable, viscosity. Lithostatic load is a function of density and depth in the deposit. The load pressure and the viscosity control the rate of loss of pore space which really is the only measure of rate of welding. We thus see that the degree of welding, in an ash flow, is dependent on three rate processes: the rates of welding, cooling, and crystallization. Crystallization is also related to the viscosity of the glass.

Of all the variables that control the cooling history of ash flows the two most important are probably emplacement temperature and the thickness of the flow or cooling unit. Because these influence all the others, or are in some way related, it is possible to interpret the observable variations in the flows in these terms. In ash flows, emplaced at approximately uniform temperature throughout, the rate of

welding will be faster at or near the base of the unit and will decrease toward the top. For this reason the transition between densely welded tuff and nonwelded tuff will invariably be much sharper in the lower part than in the upper part of a simple cooling unit. All transitions are sharper in high-temperature ash flows than in low-temperature ash flows.

In its simplest form, the welding process is the merging of viscous glass particles into a coherent mass. In very young and fresh ash flows that have not crystallized, the incipient stages of welding are marked by the first signs of coherence and brittleness in the transition from unconsolidated to obviously indurated material. In older and less fresh rocks and those which have crystallized, we must depend on visible deformation of pumice fragments or shards as a criterion for welding. From the incipient stage there is a complete gradation with progressive loss of pore space, deformation of pumice and shards, and darkening to a dense black glass. During the color change, the darkening of the pumice lumps usually precedes the darkening of the shardy matrix. The contrast emphasizes the well-known eutaxitic structure characteristic of so many welded tuffs, especially in those of low porosity (0 to 25 per cent).

The formation of a black glass appears to be the end product of the welding process; however, with aid of the microscope, an additional change becomes apparent—the gradual obliteration of vitroclastic and vesicular structures. This process begins in the collapsed pumice fragments, removing all trace of vesicular structure, and it finally extends to the shard boundaries. In some welded tuffs vesicular structure in pumice has been obliterated before the black-glass stage is reached. Elimination of all pumice structure is rare but has been noted (Ross and Smith, in press), whereas the elimination of all shard boundaries in glassy welded tuffs is not known to the writer. The elimination of all traces of shard boundaries is not achieved because of the original dusty matrix between the shards, a problem not presented by the pumice vesicles. Homogenization of all the glass shards and pumice is, then, the probable ultimate end of the welding process and will probably require an exceptionally hot, thick ash flow for its fulfillment.

Welding Temperatures Determined by Experiment

Many early workers have experimented with melting, vesiculation, and crystallization of volcanic glasses, but the first attempt to make welded tuffs by welding fragmental volcanic glass was by Boyd and Kennedy (1951, p. 327). They found that crushed rhyolitic pumice from Mono Craters, California, welded at temperatures ranging from 775° to 900° C. In later work Boyd (1957, Ph.D. Thesis, Harvard Univ.) found that the presence of water vapor effectively lowers the welding temperature. Glassy rhyolitic volcanic ash from Yellowstone Park was sealed in platinum capsules with water equivalent to 0.45 weight per cent and held for 80 hours and a pressure of 52 bars. The minimum welding temperature under these conditions was between 590° and 620° C. Increasing the time of the experiment to 2 weeks resulted in lowering of the welding temperature to between 550° and 590° C. Boyd found that the same ash, dehydrated and without addition of water in the experiment, showed slight welding at 690° C. after 72 hours at a pressure of 48 bars, and no welding at 655° C. From these and other experiments Boyd concluded that the minimum welding temperature in the presence of water vapor is about 600° C.

Smith, Friedman, and Long (1958, p. 532-533, and unpublished data) found very similar results. The method used differs from Boyd's in that the charge of unconsolidated ash was placed in a platinum-lined copper container which was in turn placed in a bomb fitted with an adjustable piston through its top. Load pressure was applied directly to the surface of the charge by the piston assemblage. A lever arm that applies load to the piston was attached to a recorder, and the rate of compaction was measured. Water pressure in the bomb was controlled externally, and the water content of the welded pellets was later measured directly or determined by analogy to quenching experiments made on crushed obsidian hydrated at different temperatures and pressures. The ash used in the welding experiments was from a nonwelded facies of the Bandelier rhyolite tuff, Valles Mountains, New Mexico.

The minimum temperature for incipient welding was below 535° C. and is controlled largely by temperature and water-vapor pressure, whereas complete welding is dependent on these variables plus load pressure and time. Time is limited by crystallization and cooling, but these factors have not yet been assessed. No satisfactory minimum temperature for complete welding (black glass) has yet been found, although it is below 635° C. at 528 p.s.i. load pressure, 300 p.s.i. water-vapor

pressure, and 50 hours time. Fifty per cent compaction was reached in one run, in about 9 days, at 535° C., 528 p.s.i. load pressure, and 150 p.s.i. water-vapor pressure. Further, for ash flows of the order of 100 feet thick or less, temperatures above 735° C. probably are necessary for the formation of a densely welded facies.

Taneda (1957, p. 99–100) reports welding experiments on the "Shirasu" (white sand), a "pumice-flow" deposit associated with Aira caldera, South Kyushu, Japan. He found that when heated in an electric furnace in the temperature range of 600° to 1200° C. in 1 to 1.5 hours it changed color from "gray white to reddish or purplish gray, a reddish tinge being deepened with raising temperatures from 600° to 1000° C." Melting began between 850° and 920° C., and at 1200° C. melting was essentially complete and the original color was restored. "Load-bearing tests of pumice" were made in the temperature range 800° to 1000° C. and at loads of 1, 3, and 5 kg/cm², the latter corresponding to 30 m depth in a "Shirasu" bed. It was concluded that the "Shirasu" would weld below 920° C. to look like "Hai-ishi" (welded tuff associated with "Shirasu") and that the nonwelded "Shirasu" could not have been emplaced above 900° C.

Because the "Shirasu" would contain little or no water at the temperatures of these experiments, and because the times were very short, it is doubtful whether or not they could apply strictly to the natural materials. Probably some welding would take place in "30 meter thick beds" at temperatures considerably below 900° C.

Compaction during Welding

Surprisingly few references have been made to the total compaction of ash flows during welding, although compaction may reach geologically significant proportions in some stratigraphic sections of densely welded tuff. The compactibility is related to the bulk porosity and therefore to the ratio of pumice to solid material. In most ash flows the initial bulk porosity must be greater than 50 per cent. This is not shown by published porosity or density curves, probably because of the difficulty of collecting and measuring representative samples of unconsolidated material, especially in older rocks.

The specific gravity and porosity curves for welded ash flows shown in Figures 5 and 6 are the only ones available. In addition to these curves Weyl (1954, p. 18, 21) gives specific-gravity figures ranging from 1.28 to 2.03 for rhyodacite tuffs from the Comalpa area and from 1.21 to 1.89 for dacitic tuffs from the Zaragoza area, El Salvador. These have not been plotted in Figure 5 because the positions of the samples were not given. The figure of 1.28 for the Comalpa tuff probably represents a bulk porosity of about 50 per cent, whereas the figure of 1.21 for the Zaragoza tuff implies over 50 per cent porosity, perhaps 55 per cent to judge from Weyl's chemical analysis. These figures seem more realistic than others published for fresh unconsolidated ash-flow materials as do the figures 0.98 (specific gravity) and 60 per cent (porosity) given by Lovering (1957, p. 1587, 1593) for the Valley of Ten Thousand Smokes "sand flow." Ross and Smith (in press) found 70 per cent porosity for part of the Battleship Rock ash flow, New Mexico. Smith, Friedman, and Long (1958) found by experiment that the compaction potential of unconsolidated material, mostly less than 4 mm in size, for the Bandelier rhyolite tuff from New Mexico is approximately 50 per cent This material was used for making synthetic "welded tuffs" and represents only the finer-grained part of the Bandelier tuff, the bulk porosity of which is considerably greater than 50 per cent.

The data indicate that where measured porosity curves show maximum porosities less than 50 per cent, they should be extrapolated to at least 50 per cent if they are to be used for studies involving thicknesses of zones or cooling units, compaction, or similar problems. Pure shards or crystals probably have a bulk porosity of about 50 per cent, and most ash flows contain pumice which commonly has 70 or more per cent pore space in rhyolitic rocks. Roughly then, an ash flow containing 50 per cent by volume of pumice or pumiceous ash of 70 per cent porosity and 50 per cent shards would have a bulk porosity of 60 per cent and a specific gravity of about 0.94 if it were 100 per cent glass, or about 0.99 if it were 50 per cent crystals of sanidine and quartz. The ratio of crystals to glass will not cause significant variation in the bulk density of unconsolidated ash, except in very mafic rocks.

Because most ash flows contain some pumiceous materials but probably not more than 50 per cent we can expect that they will have initial porosities between 50 and 60 per cent. Obviously, an ash flow having 60 per cent pore space could compact to a welded tuff having less than half the original thickness. Probably very few entire sheets have undergone such

compaction. Those in which it has occurred would have been very hot, thus more thoroughly welded, or very thick allowing a thick zone of dense welding to form, or, rarely, may have had exceptionally high initial porosity. For simple cooling units that contain densely welded tuff it is possible to get an approximate figure for the thickness before welding without density or porosity measurements, if the thicknesses of the three basic zones (*see* section on Zones and Zonal Variations) are known. Assumption of 50 per cent initial porosity probably gives the figure closest to a minimum. The calculations: 2 times the thickness of zone of dense welding plus 1½ times the thickness of zones of partial welding plus the thickness of zones of no welding equals total thickness before welding. This method is subject to some error but is a useful approximation.

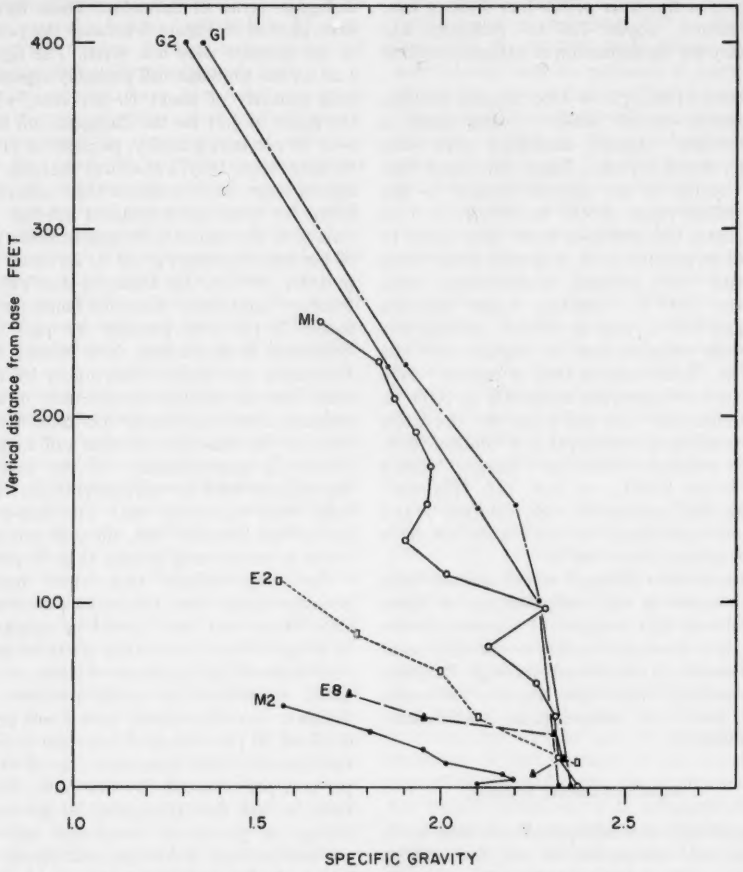

FIGURE 5.—SPECIFIC-GRAVITY CURVES FOR WELDED-TUFF SHEETS

G curves: Bishop tuff, California (Gilbert, 1938)
*M*1 curve: Motutere "ignimbrite," New Zealand (Marshall, 1935)
*M*2 curve: Arapuni "ignimbrite," New Zealand (Marshall, 1935)
E curves: Chiricahua "welded tuff," Arizona (Enlows, 1955)

Crystallization

Crystallization during cooling in pyroclastic deposits was noted by Zambonini (1919) who

described several unusual fluorine-bearing minerals from the classical Italian piperno. Fenner (1937, p. 236) found tridymite and "orthoclase" in the Valley of Ten Thousand Smokes "sand flow." Spherulitic crystallization was described by Marshall (1935, p. 347), and Mansfield and Ross (1935, p. 320) in the New Zealand ignimbrites and the Idaho welded tuffs, respectively. Marshall (1935, p. 345–350) discussed in great detail the crystallization structures and textures in the "ignimbrites" and recognized in some deposits a relationship between depth and cooling rates and the type of crystallization structures formed. Gilbert (1938, p. 1856) emphasized that crystallization in the upper part of the Bishop tuff sheet was influenced by rising gases. Westerveld (1942, p. 211–212) has described the effects of "pneumatolytic" crystallization in the Pasoemah tuffs, South Sumatra, and Toba tuffs (1947, p. 34), North Sumatra.

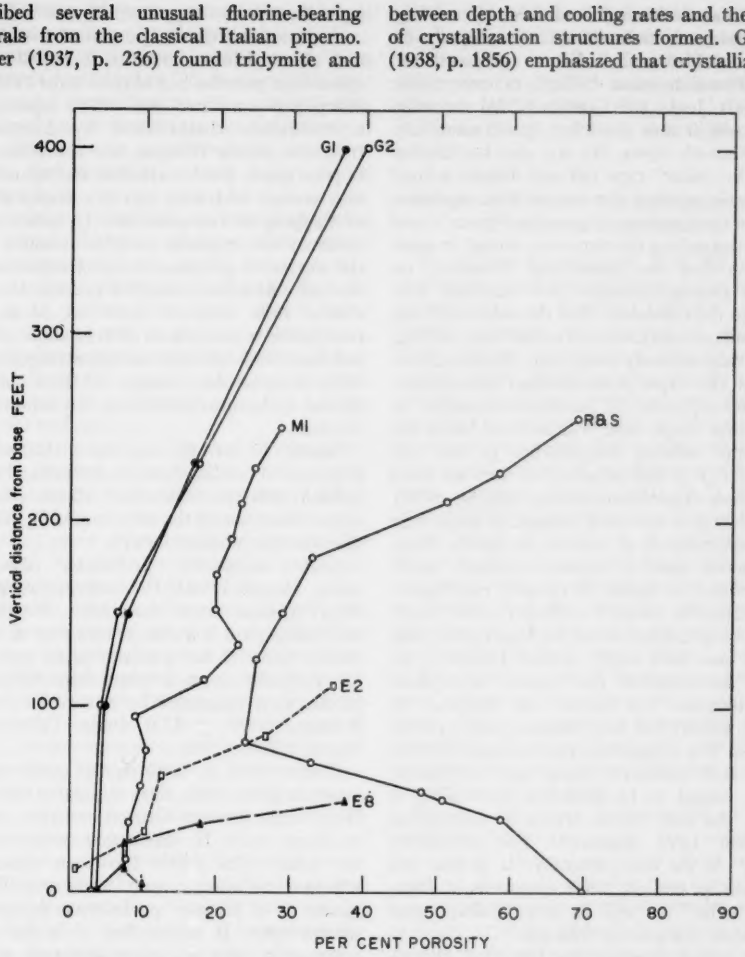

FIGURE 6.—POROSITY CURVES FOR WELDED-TUFF SHEETS
G curves: Bishop tuff, California (Gilbert, 1938)
*M*1 curve: Motutere "ignimbrite," New Zealand (Marshall, 1935)
E curves: Chiricahua "welded tuff," Arizona (Enlows, 1955)
R and *S* curve: Battleship Rock "ash flow," New Mexico (Ross and Smith, in press)

Fenner (1948, p. 883) proposed that induration in the Cordillera tuffs of Peru was due primarily to pneumatolytic "recrystallization"

rather than "softening" and union by welding, and that such tuffs should not properly be called "welded tuffs." Fenner proposed the local Peruvian name "sillar" to cover these materials. Jenks and Goldich (1956) restudied these tuffs in more detail but agreed essentially with Fenner's views. No one who has studied both the "sillar" type tuff and densely welded tuff would confuse the two in their optimum state of development or question Fenner's need for distinguishing the two, even though in some deposits they are transitional. However, we should perhaps question the reasoning that leads to the conclusion that the induration was the result of crystallization rather than welding.

We may seriously doubt that the crystallization of the vapor-phase mineral assemblages, sanidine-tridymite or sanidine-cristobalite in the "sillar" type tuffs, was *initiated* below the minimum welding temperature of the tuff (<550° C.). If this be admitted then ash flows that show crystallization must also be mildly welded if they are thick enough to show even mild deformation of pumice or shards. Some compaction could be expected without visible deformation of shards of pumice, resulting in an incipiently welded coherent rock which after crystallization would no longer show that it had once been mildly welded. However, the writer has observed that coarse vapor-phase crystallization will render an incipient to mildly welded tuff less coherent after crystallization. The comparison can be made between fresh and unaltered glassy and crystalline rocks, judged to be otherwise equivalent, in which the first visible traces of compaction folliation have appeared. The crystalline "sillar" is the more crumbly. It is also less resistant to erosion. Some specimens of Peruvian "sillar" as well as other "sillar" type tuffs show compaction foliation.

We cannot question the fact that crystallization of unconsolidated ash and pumice would result in material more coherent than it was before crystallization, but we can suggest that most completely crystalline "sillar" was a coherent, albeit porous, rock, owing to welding before it crystallized.

Ross and Smith (in press) have shown by X-ray studies that cristobalite rather than tridymite is the common form of silica, where crystallization has been confined to the glass shards or welded glass masses. Tridymite, on the other hand, is more common as discrete crystal or crystal aggregates in pore space where its crystallization is believed to be related to active vapors. Two distinct types of crystallization are suggested: (1) devitrification —characterized by intergrowths of cristobalite and alkalic feldspar, mainly in axiolitic and spherulitic growths but also in some "closed" lithophysal cavities; and (2) vapor-phase crystallization—characterized by growths of tridymite, alkalic feldspar, and other minerals in pore space. Porous axiolites and spherulites may contain tridymite, but this simply means overlapping of two processes. In general, the tendency for tridymite to form increases with the degree of porosity of the groundmass of the tuff within the crystalline portion. In some welded tuffs tridymite does not occur, yet cristobalite is common to all crystalline welded tuff sheets that have not undergone diagenetic or other metamorphic changes. In these older or altered rocks quartz replaces the other silica minerals.

Nearly all writers who have studied the products of cooling-history crystallization in welded tuffs have identified alkalic feldspar and at least one of the polymorphs of silica as the principal minerals formed.

SILICA MINERALS: Tridymite, ordinarily easily identified with the microscope, is the silica mineral most frequently mentioned, and, although it is a common mineral in many welded tuffs, it has probably been mistaken for cristobalite in many descriptions. Cristobalite has been recognized by Zavaritsky (1947), Williams (1952, p. 173), Bailey (1957), and Martin (1959, p. 404).

Enlows (1955, p. 1230) reports quartz as the common groundmass silica mineral in Member No. 6 of the Rhyolite Canyon formation, based on X-ray study. He states that tridymite was not found. The writer has been somewhat reluctant to accept most interpretations of quartz as a primary groundmass mineral in welded tuffs. It seems that it is far more common in older and more obviously altered rocks, whereas in fresh Pleistocene and Recent rocks it has not been recorded. By analogy to thick rhyolite flows it should occur in some thick sheets of welded tuff. The writer is now convinced that primary groundmass quartz should occur in any simple cooling unit of rhyolitic composition more than about 600 feet thick. The necessary thickness for the formation of quartz will no doubt vary with emplacement temperature, vapor pressure, and perhaps other factors. Enlows' Member 6 is 880 feet thick, and there seems little doubt that the quartz reported by him is primary; however, this quartz should give way at least in the upper part of the sheet to cristobalite

and perhaps tridymite, unless these minerals have been replaced by secondary quartz.

Problems of this type need a much more thorough investigation before we can arrive at any sound generalizations. The writer (Smith, in press) has proposed that the distribution of the silica minerals is zonal and is related to depth and degree of welding within simple cooling units.

FELDSPARS: Very little can be said about the groundmass feldspars in crystallized welded tuffs, and systematic studies using modern X-ray techniques have not been made. Feldspar occurs as discrete crystals or crystal aggregates with tridymite in the vapor-phase zones, as submicroscopic intergrowths with cristobalite in devitrified zones, and as granophyric intergrowths or discrete miarolitic cavity crystals with quartz.

Westerveld (1942, p. 211) and Martin (1959, p. 404) reported albite; Gilbert (1938, p. 1845) and Enlows (1955, p. 1230) reported sanidine; Fenner (1948, p. 883), orthoclase or natrosanidine; many other writers have reported "alkalic" or "alkali feldspar," "potassium feldspar," or simply "feldspar."

Probably in most silicic welded tuffs the groundmass alkalic feldspar is a form of sanidine that probably bears a relation to the bulk composition of the host rock. It is suggested that it was formed above the solvus temperature unless it be admitted that it might form metastably, as do tridymite and cristobalite. This suggestion is prompted by the fact that, in some thick cooling units, single crystals of sanidine have exsolved to crypto- or microperthites.

The feldspars also probably bear a systematic relationship to their position within the cooling units. Detailed studies of specific units may be very helpful in formulating a framework for ranges of crystallization temperatures.

Gilbert (1938, p. 1850) reported overgrowths of sanidine on sanidine phenocrysts in the upper parts of the Bishop tuff where they were presumably deposited by rising gases. Similar rims are found in some "vapor-phase zones" in the Bandelier tuff, New Mexico.

OTHER MINERALS: Minerals other than the feldspars and silica minerals that form during groundmass crystallization are not well known. Among the more common are magnetite, biotite, fayalite, and sodic amphibole. These may form deep within cooling units. Another group of near-surface and surface fumarolic minerals, including magnetite, are known largely from the work of Fenner, Allen, and Zies in the Valley of Ten Thousand Smokes, Alaska (Zies, 1929). Most of these are halides, sulfates, and sulfides of metallic elements. Magnetite has also been recorded as a fumarolic mineral by Mackin (1952, p. 1337–1338), Williams (1942, p. 86–87), and Gilbert (1938, p. 1851).

Behavior of Phenocrysts and Inclusions during Welding and Crystallization

It has not been shown that the welding process *per se* has caused changes in phenocrysts in any rocks yet described; however, crystallization processes and the rate of cooling do cause changes in individual crystals of some welded tuffs. Sanidine phenocrysts undergo changes, in optical properties and degree of unmixing to cryptoperthites, which are related to the cooling history of given vertical sections. The catalytic effects of hot vapors may facilitate these changes. Boyd (1957 Ph.D. Thesis, Harvard Univ.) has made the only quantitative X-ray study of unmixing of sanidine which shows in the Yellowstone welded tuffs a clear-cut relationship to position in the unit. In a qualitative way the unmixing phenomena in sanidine can be observed in some welded tuffs by the change from clear "glassy" sanidines in the chilled portions of the units to chatoyant ones (moonstones) in the more slowly cooled parts. This is a common phenomena in the Bandelier tuff, Valles Mountains, New Mexico, and the writer has observed it in other sanidine-bearing welded ash flows.

Whereas unmixing is detectable in some sanidine phenocrysts under the microscope, particularly those from very thick cooling units, it does not resemble the unmixing to coarse microperthite seen in the smaller, thinner sanidine crystals formed in cavities during cooling of the same unit. An excellent example illustrating this difference in degree of unmixing is a specimen the writer collected from Bonita Canyon, Chiricahua National Monument, Arizona. These rocks were described by Enlows (1955), and the specimen is from Enlows' Member 6 for which he gives a thickness of 880 feet. Chatoyant sanidine phenocrysts show only faint visible traces of unmixing, but chatoyant tabular "sanidine" crystals from miarolitic cavities are coarse microperthite.

Biotite and hornblende, commonly green to shades of brownish green in chilled glassy parts of ash flows, are often changed to deep red brown or sometimes completely destroyed in the crystalline parts of welded ash flows.

The degree of destruction seems related to the intensity of vapor-phase activity within units of given compositions, and the thickness of the cooling unit plays a major role in controlling intensity of vapor-phase activity. This deuteric or vapor-phase alteration ranges from a very minor role in some cooling units to a major one in others. Where it was intense nearly all minerals and inclusions, except possibly zircon and similar accessory minerals, show some effects, although commonly no more than border corrosion. Specifically the writer has observed varying degrees of alteration or corrosion in phenocrystic quartz, sanidine, plagioclase, amphibole, clinopyroxene, orthopyroxene, fayalite, and sphene.

No systematic studies of these alteration effects have been made, and they are another interesting field of inquiry. Excellent examples of biotite and hornblende alterations are found in welded quartz latite ash flows throughout the Basin and Range province. Some of these from southwestern Utah and adjoining parts of eastern Nevada have been studied by Mackin (1952, p. 1338, and oral communication), and by Cook (1957, p. 57).

Zones and Zonal Variations

Nearly all workers since the 1930's who have recognized the fundamental pyroclastic nature of welded tuffs have also observed and described systematic textural changes, particularly in vertical sections, within given units. Probably all significant textural variations that occur among welded tuffs have been found and described (Marshall, 1935; Mansfield and Ross, 1935; Gilbert, 1938; Enlows, 1955). Most of these can be seen in the illustrations of Ross and Smith (in press), Smith (in press), and Martin (1959). Good illustrations of marked secondary flowage features have not yet been published.

It is improbable, but not inconceivable, that all the important known textural variations resulting from welding and crystallization could be found in the same composite sheet. Most of them are possible in simple cooling units. This fact makes it difficult to classify welded tuff sheets on a purely descriptive textural basis. Most of the available descriptions are based on one or a few vertical sections which probably only rarely characterize the three-dimensional sheet. It is true that some welded tuff sheets at first glance seem to be monotonously uniform over great areas and different enough from other welded tuffs to be considered distinct types. For example, it would be improbable for the Walcott tuff (Stearns and Isotoff, 1956, p. 23) to grade laterally as the same flow unit into a crystalline tuff of the Peruvian sillar type. However, it might merge with a cooling unit which contains crystalline sillar-type material and there lose its identity as densely welded crystalline, rather than vitric, welded tuff, without the lithophysae and spherulites that now characterize it in the section at American Falls, Idaho. Thickening of the American Falls section from 25–50 feet to 75–100 feet (as a simple cooling unit) would unquestionably have resulted in the formation of a persistent crystalline zone which might predominate over the present glassy section.

In contrast to the Walcott tuff, which shows dense welding in a 25-foot thick unit, the Battleship Rock ash flow (Ross and Smith, in press) shows no dense welding in a cooling unit about 250 feet thick, but its initial porosity of about 70 per cent has been reduced to less than 25 per cent in the zone of maximum deformation, and no crystallization has taken place during cooling. The Walcott tuff, on the other hand, shows incipient crystallization in the form of spherulites and lithophysae and clearly was emplaced at the higher temperature. Clearly the evidence presented by these two sheets, together with such data as are available from other sheets, indicates a relationship between the emplacement temperature, the degree of welding, the thickness of the sheet, and the retention of the glassy state during cooling. For any given temperature of emplacement there is apparently a corresponding maximum thickness of the simple cooling unit, below which no crystallization will take place during cooling. It is not anticipated that composition, gas content, or cooling rates in different environments will, in general, show differences great enough to invalidate this generalization for gross comparisons.

In a general way, maximum vitric cooling units emplaced at high temperature are thinner, more densely welded, and show sharper transitions between nonwelded and welded material than the maximum vitric cooling units emplaced at lower temperatures. Thicknesses greater than the maximum vitric unit for any given temperature will result in crystallization, incipient at first (Walcott tuff of American Falls, Idaho) but persistent as a continuous zone of crystallization at greater thicknesses. This zone exhibits an increasing number of facies as the unit thickens further. These different facies or zones of crystallization are

related to the degree of welding of the tuff on which they are superimposed as well as to their positions within the units and to other factors.

Smith (in press) has attempted to show the general relations between thickness, temperature, and the formation of textural and structural zones by means of hypothetical diagrams. Three basic zones are recognized: (1) the zone of no welding, (2) the zone of partial welding, and (3) the zone of dense welding. Where dense welding occurs in a unit, the other two zones have upper and lower counterparts which merge with lateral thinning or near the borders of the sheet. The lower zones of no welding and partial welding are thinner than the upper ones in simple cooling units and may be locally absent owing to special or extreme conditions of temperature and pressure. Crystallization superimposed on the basic zones may be recognized as (1) the zone of devitrification, (2) the zone of vapor-phase crystallization, (3) the zone of granophyric crystallization, and (4) the zone of fumarolic alteration.

This scheme of representation is largely qualitative, but, with the accumulation of reliable measurements of the various zone thicknesses and positions within a variety of cooling units, it may develop into a quantitative scheme leading to the solution of problems of heat loss during emplacement, emplacement temperatures, rates of cooling, direction of source, and many others. In future studies geologists should, if possible, record the zonal characteristics, especially of complete cooling-unit sections, quantitatively. The zonal patterns of simple cooling units must be understood before we can hope to evaluate compound cooling.

Deposits Emplaced at Low Temperature

Ash flows probably can be emplaced at any temperature up to a maximum eruption temperature, perhaps of the order of 1000° C. If the minimum welding temperature is taken to be about 500° C., there is a wide temperature range through which endless variations can occur.

We are not yet able to correlate specific ash flows with specific temperatures, but a gross relative scale is possible. Because thickness is a most important factor in controlling degree of welding, many thin ash flows emplaced above their potential minimum welding temperature do not show welding. These are ordinarily not distinguishable from deposits emplaced below their minimum welding temperature, although the presence of a red zone caused by oxidation of iron probably indicates a higher temperature. Ferromagnetic properties may be useful in distinguishing nonwelded ash flows emplaced above the minimum welding temperature which is close to the Curie point.

The designations high, low, and intermediate must be arbitrary and are only expedient subdivisions for ordered discussion.

Low-temperature deposits, then, are those in which little or no welding or crystallization is apparent in thick cooling units. These include deposits such as the Crater Lake and Krakatoan pumice deposits, the "Shirasu", and many others. Because some ash-flow fields contain units emplaced at high, low, and intermediate temperatures (Pl. 2, fig. 2), but presumably were derived from the same magma and erupted at the same temperature, the low-temperature deposits offer the best evidence for a cooling mechanism before the ash flow is formed. (*See* section on Vertical Component.)

Deposits Emplaced at High Temperature

Perhaps the best examples of high-temperature deposits are those from Idaho (Mansfield and Ross, 1935; Stearns and Isotoff, 1956). These rocks occur in sheets ranging from 15 to 50 feet thick and contain a zone of densely welded tuff to within a few feet of the surface and locally a zone of fused air-fall tuff at the base. These facts attest to their high temperature of emplacement. A high magmatic temperature is suggested by the paucity of phenocrysts. Probably these materials were derived from very hot and very dry magmas. Obsidian formed in the welded tuff from the Ammon quadrangle (Mansfield and Ross, 1935, p. 319–321; Ross and Smith, in press) has a water content of 0.15 per cent, similar to many normal obsidians of rhyolite flow or dome origin. The Walcott tuff of Stearns and Isotoff (1956), considered by Mansfield and Ross (1935, p. 310) to be equivalent to the tuffs studied by them, has been restudied by H. A. Powers (oral communication), who reports water values for the obsidian facies as low as 0.06 per cent. Water contents of this order are probably the lowest known for volcanic glasses and give further indication of the high emplacement temperature if not the low-water content of the magma.

Mansfield and Ross (1935, p. 312) considered that the tuffs described by them had been

emplaced by a great air-borne cloud of incandescent particles and cited as evidence the fact that the tuff occurred at different elevations, without significant change in thickness, over a vertical relief of over 2500 feet. The distribution of the tuff suggested that it had been sprayed or "ducoed" over the irregular surface. This connotation has been widely cited in the literature and should be re-evaluated. C. S. Ross (oral communication) no longer holds this view; he now thinks these tuffs were formed from normal ash flows, and that their distribution may be explained by post-emplacement deformation.

Because these welded tuffs are an important link in the development of present-day concepts, they should be restudied. The possibility that, as normal ash flows, they were deposited with very low initial dips, probably less than 2°, might be critical to the interpretation of the geology of the very large area covered by these tuffs. Mansfield and Ross (1935, p. 311) have suggested that this might be of the order of 3000–5000 square miles. The very steep dip of 24° shown by them in one of Mansfield's (1935, p. 310) stratigraphic sections seems incompatible with a primary dip for a welded tuff sheet and suggests that at least part of the sheet has been deformed.

Deposits Emplaced at Intermediate Temperatures

Most welded ash flows are in the intermediate-temperature category. Depending on thickness, there are all degrees of welding and crystallization within this group of rocks. Thin cooling units look like low-temperature deposits, and parts of thick cooling units are not distinguishable from high-temperature deposits. Perhaps the key to relative emplacement temperature is the thickness of the upper zone of partial welding in simple cooling units. This should vary systematically between low and high temperature, but for comparisons between two units it must be equated with initial emplacement thickness or expressed as a ratio to the collapsed thickness.

On the low-temperature side of the intermediate-temperature category are the thick vitric cooling units that show moderate welding, whereas on the high-temperature side are the relatively thin cooling units that show crystallization, a zone of dense welding, and probably a weakly developed and thin vapor-phase zone. The vapor-phase zone, including the crystalline "sillars," probably forms in an optimum temperature range within this intermediate category.

The vapor-phase zone requires a porous matrix for its formation (Pl. 1). At increasingly higher temperatures, the zone of dense welding would probably thicken, whereas the porous upper zones would thin, until they become a part of the "chilled" vitric mantle of the sheet, leaving no space for the formation of the vapor-phase zone. In these higher-temperature rocks the vapor-phase zone would probably give way to a zone of lithophysae in thin units or to a zone of miarolitic lenticles in thick units. Composition, however, becomes important to the problem at this point, and, in general, more basic deposits show less crystallization throughout than do the rhyolites.

Deposits that Record a Change in Temperature

Some ash-flow fields, perhaps many, record variations in emplacement temperature as shown by the different degrees of welding and of crystallization of individual flows or cooling units of equivalent thicknesses. The indicated changes in temperature are particularly significant in areas where all units have a common source. The change may be recognized within a cooling unit or in two or more units. Changes found by the writer appear to be, for the most part, systematic, indicating a general increase in emplacement temperature in successive flows.

Weyl (1954, p. 17–23) has described dacitic and rhyolitic welded tuff deposits from the Zaragoza and the Compalpa regions of El Salvador, respectively. They are of the order of 100 m thick and decrease in density from the top to the bottom. Weyl has noted the difference between these and other welded tuff sheets which increase in density toward their base. Weyl (1954, p. 27) suggests that the earlier "nuées ardentes" brought forth relatively cool material, whereas the later ones carried progressively hotter material that was more susceptible to welding.

We cannot tell, from Weyl's description, the specific type or types of units represented by either the Comalpa section which is 100 m thick or the Zaragoza section which ranges from 30 to 100 m, but the inference is that the changes he noted from top to bottom of both sections are continuous and progressive. The Comalpa section shows a range in specific gravity from 2.03 near the top to 1.28 at the base, accompanied by a change from obvious welding at the top to no welding at the base.

Systematic changes of the type described by Weyl may be explained in several ways depending on the kind of units involved. If the changes are related to differences in emplacement temperatures, they may represent a change in magma temperature as progressively lower levels of the magma chamber are drained during the eruption, or they may represent progressively more efficient conservation of heat during the formative stages of successive ash flows.

The detection of evidence for changes in emplacement temperature in successive units is difficult in many deposits. Variable thicknesses of flow units, the order in which they are stacked, and the length of time between successive flows can create compound cooling units in which systematic emplacement-temperature changes are not obvious and in which apparent changes are not real. Deposits may eventually be found that show an increase in density and degree of welding from top to bottom, but which were formed by stacking of successively hotter thinner flows.

Deposits that Record a Change in Composition

Deposits from a common source that record a change in composition in successive ash flows may be the most significant of all deposits geologically. These changes are of three kinds: (1) change in magma chemistry; (2) change in crystal–glass ratio; and (3) change in type of cognate vesicular fragments. A change in magma chemistry is usually accompanied by the physical changes of (2) and (3) and may be recorded in one compound cooling unit.

The Crater Lake field is probably the best example of a change in magma chemistry yet described. Williams (1942, p. 98) states that the total volume of pyroclastic material laid down during the culminating eruptions of Mt. Mazama is between 9.75 and 11.75 cubic miles. Six to 8 cubic miles of this consists of "glowing-avalanches" deposits, seven-eighths of which is dacite pumice and ash; the rest is crystal-rich "basic scoria." The silica content of the dacite pumic ranges from 66.4 to 68.9 per cent; that of the "basic scoria" from 53.9 to 56.9 per cent. There are no intermediate andesitic ejecta. Mineralogically the "basic scoria" differs from the dacite pumice in having more ferromagnesian minerals, hornblende and olivine, and more calcic feldspar, and dark-brown glass. It also contains more lithic inclusions. Most remarkable is the fact that the transition between the underlying "pumice flows" and the later "scoria flows" takes place within a few feet (Williams, 1942, p. 83). The transition was sudden, without a detectable time break, and it must be concluded that there was an essentially continuous evolution of material. Locally the "scoria flows" have well-developed columnar structure; the "pumice flows" do not.

Commonly volcanoes that have erupted silicic magma later erupt more basic magma. In most instances we cannot evaluate the changes that took place between eruptions. At Crater Lake there was no time break, and the suddenness of the transition leads us to consider possible conditions in the magma. The continuity of the eruptions that produced the dacite pumice and "basic scoria" suggests they came from a single magma chamber where they must have existed as magmas with a sharp interface. The frequency of formation of simple calderas, like Crater Lake, in composite cones supports the single-chamber hypothesis.

The Crater Lake deposits are probably not unique. Contrasting magmas of rhyolitic and basaltic compositions are known from air-fall pyroclastic eruptions in Iceland. These are very small-volume deposits. Hekla (Thorarinsson, 1954, p. 32–45) is a notable example. In contrast to these the volume (order 5, Fig. 3) of the Crater Lake ejecta is large. The volume of pyroclastic ejecta making up the Bandelier tuff, Valles Mountains, New Mexico, is approximately 40–50 cubic miles (order 6, Fig. 3). The Bandelier ash flows record a change from early high-silica rhyolites to quartz latites at the end of the last ash-flow cycle.

The lavas that preceded the ash flows include quartz latite, rhyodacite, and more basic types.

Even the very large volume (order 7, Fig. 3) deposits may show similar trends. For example, van Bemmelen (1949, p. 694) states that the Toba eruption was preceded and succeeded by andesitic volcanism. The outburst of "granitic magma" was only an interlude in the continuous volcanic activity that normally produced "dioritic" magmas. Van Bemmelen (1949, p. 691) says that the Toba tuffs were probably produced by one, or a short succession of, cataclysmic outbursts, because no distinct stratification has been observed throughout most of the stratigraphic section. Some layering, however, is observed in the higher parts. Is it possible that a change in compositon is recorded?

Both andesitic and rhyolitic magmas were involved in the formation of the "glowing-

avalanche" deposits of the Sudbury basin (Williams, 1957, p. 69); andesitic "avalanches" followed rhyolitic eruptions.

Some ash-flow fields, composite sheets, and cooling units record a change in composition in their late stages. The frequency of such occurrences cannot be determined from published information; however, where such changes occur in young and fresh rocks, they offer an unprecedented approach to the study of problems of geochemical equilibrium in silicic magma chambers and will perhaps provide new information on the behavior of magmas. Time intervals, with their attendant possible changes, are drastically reduced or eliminated, a condition not shared by normal volcanic or intrusive sequences.

Origin of the Magmas

The evidence suggests that welded ash flows are not unique to any type of vent, source area, or composition of magma. Nevertheless they appear to be formed largely from fissures, are seemingly more common among large-volume deposits associated with subsidence structures, and more abundant among silicic magmas.

The large-volume deposits are confined to continental mountain belts or to wide island-arc systems, or in short to those environments that seem most likely to produce large volumes of granitic magma. This brings us directly to the contrasting concepts of the separateness or unity of volcanic and plutonic associations discussed by Kennedy and Anderson (1938).

Deposits of small volume, whatever their ultimate origin, have had an immediate origin within a volcanic dome or cone. Larger volumes related to caldera collapse were almost surely derived from shallow magma reservoirs having areal dimensions at least as large as the calderas, which in order 5 of Figure 3 range up to at least 20 square miles (Crater Lake). At depth these sites would logically be occupied by small plutons, and most are associated on the surface with more mafic volcanic rocks both older and younger than the ash-flow eruptions.

Some of the source structures associated with fields of order 5, 6, and probably 7 (Fig. 3) are true cauldron-subsidence structures, complete with ring volcanoes and a central stock; the latter is inferred from an observable uplifted central dome of pre-subsidence rocks.

Oftedahl (1957) has recently shown that the volcanic rocks of the ring structures near Oslo, Norway, long thought to be lava flows and intrusive rocks, are really welded tuffs and related effusive rocks. Buddington (1959) has recently reviewed the granite problem and has called attention to the clear-cut, volcanic-plutonic relations among the epizonal plutons. Buddington (1959, p. 680) cites Billings (1943, p. 134) as having found that 11 out of 30 ring complexes reported at the time of his survey contain volcanic rocks. Probably all of these should be re-examined for welded tuffs.

The ring complexes, then, offer the best proof of a direct relationship between extrusive and plutonic rocks. Within some of these structures, ash-flow eruptions must have come directly from magma bodies that later congealed as plutons. The ring-dike complexes are commonly grouped in such a way as to suggest that they are only the upper part of a large subjacent complex (Buddington, 1959, p. 680-690; Jacobson, MacLeod, and Black, 1958).

The calderas associated with deposits of order 6 (Fig. 3) have areas of the order of 90 to 150 square miles; presumably the underlying magma bodies are at least as large and probably larger. Many ring structures are of comparable size although some ring-dike complexes are notably larger. Billings (1928, p. 86) says that the White Mountain New Hampshire "batholith" and associated volcanic rocks cover an area of 680 square miles.

Deposits of order 7 (Fig. 3) were almost necessarily derived from even larger magma bodies. The Toba volcano-tectonic depression occupies an area of approximately 1200 square miles which suggests a magma body of at least 60 miles long by about 20 miles wide or equal to the dimensions of the structure. Van Bemmelen (1939, p. 130) states that the estimated volume of pyroclastic rocks erupted approximates the volume of the Toba subsidence.

Buddington (1959, p. 681) has calculated that the 2300 cubic miles of rhyolitic volcanic rocks belonging to the Potosi volcanic series of the San Juan Mountains of Colorado (Larsen and Cross, 1956, p. 94) are equivalent to a "granite batholith" 230 square miles in area and 10 miles deep.

Certainly volcanic eruptions have produced volumes of pyroclastic rocks, principally in the form of ash flows, of enormous proportions, over a short period of time. It is inferred that greater volumes of magma were not erupted, but remained behind to congeal as large plutons. The present evidence, from known volcanic fields and from plutonic complexes that contain subsided blocks of volcanic rocks, indicates

that the immediate sources of many ash-flow fields are small to large granitic to dioritic complexes; they are not, however, directly traceable to the deeper-seated great batholithic complexes. Nevertheless the tectonic settings and regional grouping of some ash-flow fields suggests common origins from deeper-seated bodies.

Considering now the immense deposits of areas like those of the Basin and Range province of western United States, perhaps the important question is whether or not the ash-flow fields truly represent the magmas from which they were immediately derived. In other words, are the rhyolite ash-flow fields derived from granite magma or do they represent only the differentiated part of bodies of quartz monzonite or granodiorite? If they represent only the differentiated part, are they all or only part of that fraction? At Crater Lake the evidence shows that an ash-flow eruption can drain the entire silicic differentiate, at least on a small scale. Could this not also happen on a large scale? Quartz latite and rhyodacite ash-flow sheets are common, in addition to the rhyolite ash flows, throughout the Basin and Range province. Are these separate entities from other source areas or are they related genetically to the rhyolites, perhaps in a systematic succession? The answers to these questions may have real significance to students of ore deposits. It is suggested that the answers may be found by careful studies of ash-flow successions from common source areas.

Geochemistry

Valley of Ten Thousand Smokes

The studies of Shipley (1920), Sayre and Hagelbarger (1919), Allen and Zies (1923), and Zies (1924; 1929) in the Valley of Ten Thousand Smokes contain nearly all the known quantitative information on the geochemistry of ash flows. The essentials of these studies are now well known and include the tremendous amount of chlorine and fluorine given off by the fumaroles; the transport in the vapor phase of a large number of metallic elements and their deposition in sublimate minerals around the mouths of fumaroles and in fissures; the detection of most of these same metallic elements in the unaltered glassy pumice; and their release from the pumice by attack of steam and acid gases. The reader is referred to Zies's (1924; 1929) papers for details. These studies were among the finest of their kind, and their importance to students of ore deposits is well recognized. Especially important is the positive evidence for low-concentration, vapor-phase transport of metallic elements from seemingly ordinary magmas. Unfortunately, for students of pure magma chemistry, these deposits are formed from a mixed magma of andesite and rhyolite (Fenner, 1923, p. 38; Williams, Curtis and Juhle, 1956, p. 129).

The controversy over the source for the fumaroles has not yet been resolved. Griggs (1918, p. 97), Fenner (1923, p. 43), Allen and Zies (1923, p. 95), and Zies (1929, p. 4) have debated this point; although they recognize that some of the smaller fumaroles may have originated in the ash flow itself, they concluded that most fumaroles had a deeper-seated source.

Examination of the interiors of dissected prehistoric ash-flow deposits indicate that during cooling of many deposits vapor-phase activity was common. Moreover, a large proportion of ash flows, in particular those that crystallized during cooling, gave off vapors at their surface for many years.

This geologic evidence supports the belief that the "sand flow" generated a large proportion of its own fumaroles with some help from surface waters converted into steam by contact with the hot ash. Indeed, Williams (1954, p. 58) says, "Fumarole cracks gradually disappear downward within the avalance deposits; hence the ore-minerals were sublimated by gases given off by the fragmental ejecta and not by a deeper body of magma." The data published by the early workers at the Valley of Ten Thousand Smokes, however, suggest that many of the fumaroles rising from the "sand flow" in the vicinity of Novarupta, Falling, Broken, and Cerberus mountains are related to, or contaminated by, a deeper source. Allen and Zies (1923, p. 105) say, "The hottest areas (in 1918) are in the central part of the main Valley opposite Knife Creek Valley, in the Novarupta Basin, in Broken Mountain Valley and at a few points at either end of Knife Creek Valley. They all fall in a nearly circular but not continuous zone, with a salient extending out from Broken Mountain into the middle Valley." All the incrustations analyzed by Zies (1929, p. 41) containing more than 0.1 per cent boron were found in this same area, and fumarolic activity persisted here locally at least as late as 1952 (Muller and others, 1954, p. 66).

Lovering (1957, p. 1587–1588) has discussed the evidence (from the earlier papers of Shipley, 1920; Allen and Zies, 1923; and Zies, 1929) bearing on the geographic control of sulfur-acid

gases and carbon dioxide and concludes that these too existed in higher concentrations in the same general area mentioned above. These facts indicate that not all of the "Ten Thousand Smokes" were from the "sand flow", and we may infer that an area of about a 3- or 4-mile radius centering on Novarupta was subject, in addition to "sand-flow" gases, to gases from a source beneath the ash flow. Contamination of ash-flow fumaroles farther down the valley by ground-water transport of the deeper-seated materials seems a distinct possibility.

On the other hand, Lovering's (1957, p. 1587) conclusion that some source of acid gas was tapped in addition to that originally "imprisoned" in the ash flow seems based, in part, on the assumption that all the gas given off by the ash flow itself was present in void space at the time of emplacement of the ash. A more voluminous source of HF and HCl must have been the gas in solution in the glass fraction of the ash flow—conservatively, about 80 per cent of the solid material of the flow. A large proportion of the HF and HCl may have been released on crystallization of the glass. Some crystallization had taken place by 1919 (Fenner, 1937, p. 236). We may presume that crystallization was continuing in 1919. It may well be that this was the major source of the HCl and HF reported by Allen and Zies (1923, p. 143) and by Zies (1929, p. 61–68); it could have exceeded by several or many times the *volume* of free gas that could be entrapped in the voids at the high temperature of emplacement of the ash flow.

Most of the gas released during initial vesiculation of the magma was certainly lost before the ash flow formed, and most of the rest was lost before final emplacement or during or shortly after the initial stages of compaction and welding, *before crystallization began.* Great quantities of the juvenile gases coming from the "sand flow" itself, 5–6 years after the eruption were probably being released by crystallization of the glass and had little physical relation to the entrapped gas released by vesiculation.

The rate of crystallization in volcanic glass is not well known, but temperature, composition, volatile content, and time are important variables. Some ash-flow cooling units 250–300 feet thick show marked but not dense welding and have remained vitric after cooling. Others may be only 50 feet thick, yet show dense welding and have a persistent crystalline zone. Clearly the latter had a shorter cooling history than the former in spite of its higher initial temperature. The "sand flow" over most of its area seems to fit somewhere between the above examples both in thickness and in crystallinity, as well as presumably, in time for cooling. The maximum thickness of the "sand flow" is not known, but Zies (1929, p. 4) states that it is at least 100 feet thick, and this was the thickness used for volume calculation. At least mild welding and crystallization are indicated from Fenner's (1923; 1937, p. 236) allusions to indurated material, reactions with inclusions, concavity of the surface of the flow, presence of tridymite and sanidine in some of the materials, and from illustrations in Griggs's (1922) and Fenner's (1923) papers, especially those showing crude columnar structure and fallen blocks of coherent material.

In essence, the evidence forcefully implies that the time of release of gas by crystallization varies considerably in different deposits and even in the same deposit where cooling gradients differ. The "sand flow" probably underwent a period of dormancy, for release of juvenile gas, between the time of welding and compaction when most of the trapped gas was released and the time the glass began to crystallize. This time could have been 2–3 years during which fumarolic activity, if any, may have been largely supported by reheated ground water. The validity of this reasoning depends on whether or not the glass of the "sand flow" could remain oversaturated in volatiles from the time of its eruption and throughout the period ending with crystallization. In all probability it could not. However, Kozu (1934, p. 164–165) reports that, of the two types of pumice from the 1929 eruption of Komagatake, the "pumice-flow" type retained more HCl and gave it off at lower temperature (when reheated) than the air-fall type did. These specimens presumably were collected within the first year after the eruption.

The "sand flow" was formed in 1912, discovered in 1915, and the first studies of temperature and fumarolic products were during the period 1917–1919.

If the line of reasoning leading to the suggestion of a period of "dormancy" during the early years of the "sand flow" has merit, then extrapolation of amounts of volatiles given off, beyond the period of observation of the fumaroles, must be severely restricted.

Ancient Deposits

Many of us have been fascinated with the thought of finding fumarolic mineral deposits,

similar to those described from the "sand flow", in ancient ash-flow deposits. Traces of former fumarolic activity in the tops of some sheets (Mackin, 1952, p. 1337–1338 and oral communication; Gilbert, 1938, p. 1851–1854; Williams, 1942, p. 86–87; 1952, p. 156) and thick zones of vapor-phase crystallization in others make it seem likely that many ancient sheets and the "sand flow" had parallel histories. Further, many ancient deposits show evidence of having had far more intense vapor-phase activity than did the "sand flow."

Zies (1929, p. 18–19) has emphasized the ephemeral nature of the "sand-flow" sublimate species and has stressed the importance of falling temperature and steam dynamics. Especially important is the absence of magnetite and other minerals when the temperature of a fumarole falls to the point where steam condenses. The minerals are then dissolved by halogen acids and water. The persistence of such deposits can probably be ruled out for simple cooling units on the assumption that they would have conformed closely to the pattern set by the "sand flow."

Compound cooling units in ancient deposits are common. This leads us to wonder what would happen to the fumaroles and their products in a sheet at the stage of cooling of the "sand flow" during the period 1917–1919 if another ash flow were emplaced on top of it. Perhaps much would depend on the characteristics of the second flow, which, if it were hot and quickly welded, might effectively seal in the fumaroles in the lower flow. Ultimately, the deciding factors would probably be ground water and permeability. The suggestion is made that these compound cooling units that show evidence of marked breaks in their cooling history should be examined very carefully for fossil fumaroles and traces of their former products.

Many chemical analyses of rocks now known to be of ash-flow origin have been published. They record compositions ranging from basalt to rhyolite and trachyte, but show no notable differences as a group from normal lavas of similar compositions. The same statement applies, in so far as is known at present, to trace elements.

Jenks and Goldich (1956, p. 168–169) found that the "salmon sillar" from the Arequipa region, Peru, contained small amounts of soluble salts (chloride and sulfate), and halite was determined by X ray. They attributed these to introduction by "ground water, which is locally saline owing to leaching of volcanic deposits." This study suggests examination of vapor-phase zones in ancient deposits for water soluble salts, if suitable deposits could be found, located above present and past water tables.

Systematic geochemical studies of welded ash flows are needed. We are now in a position to make such studies of *limited* scope. They should be made, in the beginning, on the youngest possible fresh rocks. Probably some sheets will be essentially homogeneous chemically, and others not. We need to know the differences and their orders of magnitude.

Geochemical studies must consider the degrees of welding and types and degrees of crystallization. Major and minor elements should be determined in relation to the zonation in simple cooling units, preferably in single flows. Great care must be taken to evaluate contamination by inclusions, and in the selection of suitable samples. J. H. Mackin (oral communication, 1957) has arrived at similar ideas based on his preliminary work in southeastern Nevada.

Concluding Remarks

The problem of ash flows is in a formative stage, and a great deal of study is needed, especially to supply quantitative data. Rocks and rock bodies of ash-flow origin are now known to be so common that their recognition is no longer surprising, as it was only a decade ago. It no longer seems necessary to recapitulate the arguments or early history of development of concepts.

The importance now of recognizing these rocks, aside from pure knowledge, is because they are useful tools for the understanding of other geologic processes. Here their usefulness may be many fold, but much will depend on the accumulation of quantitative information from entire pyroclastic fields, composite sheets, or cooling units.

The important use of widespread ash-flow deposits for correlation of stratigraphic sequences, although recognized 20 years ago by Anderson and Russell (1939), is now being more fully explored and holds great promise. Many, if not all, ash-flows fields are associated with extensive air-fall beds. In and adjacent to the post-Cretaceous continental orogenic systems, we should be able to establish eventually a "tephrachronology" for the entire Cenozoic that may transcend anything envisioned today.

Detailed studies of the cooling characteristics of simple ash-flow sheets and the application

of the information to more complex sheets and to entire fields could eventually lead to concepts of eruption intervals, depositional time spans of certain rock units, and other time concepts that we do not now have.

Some of these great sheets surely represent magmas which, if intrusive, would have produced, or caused production of certain types of ore deposits. They, together with associated air-fall pyroclastic rocks, represent perhaps the closest possible approach to arrested equilibria in silicate melts that it is possible to attain in nature. In effect they may represent a quenched inverted sample of differentiating silicate magma throughout a depth of at least the upper several thousand feet in many magma chambers. To be sure most of their volatile constituents have been lost, especially those first released, but others are retained in glass, or their passage is recorded in crystallized welded tuffs. The phenocrysts remain, however, and detailed geochemical and petrologic studies of them and their relationship to their glass matrix should be made. The significance of such studies will depend on the extent to which they are guided by careful investigations of the pyroclastic fields in their entirety and should eventually encompass a variety of petrographic provinces, compositions, and fields of different volumes. Presumably the most useful data will come from those materials that represent the widest variation in composition within a single eruption cycle. All these pyroclastic deposits probably record some kind of differentiation, ranging from the extremes of a basalt–rhyolite relationship to changes in the crystal–liquid chemistry of crystallizing granite magmas.

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PALEOTECTONIC EVOLUTION OF THE CENTRAL AND WESTERN ALPS

BY RUDOLF TRÜMPY

ABSTRACT

This paper deals with the general features of Mesozoic and Tertiary rock sequences and paleogeography in the Alps. It seeks to outline the paleotectonic significance of the rocks and to discuss the structural evolution of the Alpine geosyncline up to the main deformation, with special reference to the sector between the rivers Rhine and Durance. Accent is placed on the relative independence of Alpine structures involving the pre-Triassic basement rocks and of cover nappes consisting only of Mesozoic and Tertiary formations.

Normal shallow-water deposits of platform or miogeosynclinal type were laid down over the whole area before eugeosynclinal conditions set in. The typical eugeosynclinal sediments in the central, Penninic belt of the Alps are the Schistes lustrés and Bündnerschiefer, with sills and submarine lava flows of basic volcanic rocks (ophiolites). Before metamorphism they consisted mainly of shales and of impure arenaceous and argillaceous limestones. The bathymetric environment of radiolarian cherts and associated rocks is examined, and their deep-water origin is upheld for the Alpine occurrences. Marine polygenic breccias are characteristic of geosynclinal slopes (commonly fault scarps) and not of a particular depth zone. The Alpine Flysch is a particularly significant sediment. Flysch is a thick marine deposit of predominantly detrital rocks, in part turbidites, generally without volcanic rocks, and laid down during compressional deformation of the geosyncline. Of the many different kinds of Flysch some represent transitions to either Bündnerschiefer or Molasse. Essential differences between the early synorogenic Flysch and the later synorogenic to post-orogenic Molasse are listed.

Argand's stimulating embryotectonic theory of the evolution of the geosyncline is outlined and rejected. The Triassic corresponds to a neutral interval, between the Hercynian and Alpine cycles. Early Alpine geosynclinal history was characterized by vertical or tensional movements along normal faults that limited narrow platforms and rapidly subsiding troughs. This tensional deformation weakened in the Late Jurassic, whereupon bottom relief diminished, and "para-oceanic" conditions prevailed over a large part of the Mediterranean realm. New linear welts of compressional origin arose during the Cretaceous and developed into steep island chains limiting the Flysch basins. Gradually the central part of the Alps rose above sea level, and the geosyncline migrated to the north and west. Here it was finally filled and gave way to the foreland trough in which the Molasse was laid down.

There is no simple, direct, genetic and space relationship between Hercynian structures, fault-bordered platforms and troughs of the earlier geosynclinal phase, island chains and basins of the Flysch phase, and the later nappe structures. Paleogeographical features are commonly short-lived and migratory.

RÉSUMÉ

Ce travail est consacré aux caractères généraux des dépôts mésozoïques et nummulitiques des Alpes. On a essayé d'esquisser la signification paléotectonique des divers types de formations, et de discuter l'évolution structurale du géosynclinal alpin jusqu'à l'époque des plissements principaux. La plupart des exemples ont été choisis dans le secteur entre le Rhin et la Durance.

Un bref résumé est destiné à donner les grandes lignes de la tectonique et de l'histoire géologique des Alpes. On a insisté sur l'indépendance relative des structures affectant le socle pré-triasique et des nappes de décollement.

Des dépôts néritiques normaux, caractérisant des plateformes ou des miogéosynclinaux, ont été formés partout avant l'avènement des conditions eugéosynclinales. Les sédiments caractéristiques de la zone centrale, eugéosynclinale, des Alpes (domaine Pennique s. l.) sont les Schistes lustrés et les Bündnerschiefer, avec des filons-couche et des coulées sous-marines de roches basiques (ophiolites). Avant le métamorphisme, ils étaient surtout formés de schistes argileux et marneux et de calcaires impurs, sableux et marneux. Le

milieu bathymétrique des radiolarites et des roches accompagnantes est examiné, et l'on conclut à leur origine en mer profonde. Des brèches sous-marines polygéniques sont caractéristiques des talus géosynclinaux, souvent dûs à des failles, mais non pas d'une zone bathymétrique déterminée. Enfin, il est essayé de donner les caractères essentiels des flyschs des Alpes. Ce sont des formations épaisses, marines, à prédominance détritique, et renfermant des dépôts de courants de turbidité. Il n'y a généralement pas de roches volcaniques. Le flysch s'est déposé pendant que les efforts de compression tangentielle affectaient déjà le géosynclinal. Il y a de nombreuses espèces de flysch, ainsi que des termes de transition soit vers les Bündnerschiefer, soit vers la molasse. Les principales différences entre flysch et molasse sont indiquées.

La seconde partie du travail traite de l'évolution du géosynclinal. La théorie d'Argand sur la tectonique embryonnaire est exposée; elle doit probablement être abandonée. Le Trias correspond à une période neutre, entre les cycles Hercynien et Alpin. Pendant une première phase de l'histoire géosynclinale, il y avait surtout des mouvements verticaux ou d'extension, le long de failles normales limitant des sillons subsidents et des plateformes assez étroites. Les déformations de ce type devinrent plus faibles au Jurassique supérieur. Pendant un certain temps, le relief sous-marin fut peu prononcé, et des conditions "para-océaniques" régnèrent sur une grande partie des pays méditerranéens. De nouvelles rides, dues aux forces tangentielles naquirent pendant le Crétacé, et devinrent les "cordillères", des guirlandes d'îles au relief très accentué, limitant les bassins où se déposait le flysch. Peu à peu, la partie centrale des Alpes s'éleva hors des eaux; le géosynclinal fut déplacé vers le nord et l'ouest, où il fut enfin comblé et passa à l'avant-fosse molassique.

Il n'y a pas de relation simple et directe, de nature topographique et génétique, entre les structures hercyniennes, les plateformes et les sillons de la première phase géosynclinale, les "cordillères" et les bassins de la phase du flysch, et les nappes actuelles. Les éléments paléogéographiques sont souvent assez éphémères, et sujets à des migrations répétées.

Zusammenfassung

Diese Arbeit behandelt die wichtigsten Züge der mesozoischen und alttertiären Sedimentbildungen in den Alpen. Es wird versucht, die palaeotektonische Bedeutung der verschiedenen Gesteinsgruppen zu umreissen und die strukturelle Entwicklung der alpinen Geosynklinale bis zu den Hauptfaltungsphasen zu verfolgen. Die meisten Beispiele sind dem Abschnitt zwischen dem Rhein und der Durance entnommen.

Über die Tektonik und die geologische Geschichte der Alpen orientiert ein kurzer Überblick. Besonders betont wird die relative Unabhängigkeit der Strukturen, in welche das vortriadische Grundgebirge einbezogen ist, und der nur aus mesozoischen und alttertiären Gesteinen bestehenden Abscheerungsdecken.

Normale Seichtwasserablagerungen von Plattform- oder miogeosynklinalem Charakter wurden überall gebildet, bevor die eigentliche eugeosynklinale Entwicklung einsetzte. Die charakteristischen eugeosynklinalen Sedimente des zentralen, penninischen Faziesbereichs sind die Schistes lustrés und Bündnerschiefer, mit Lagergängen und Ergüssen basischer Gesteine (Ophiolithe). Vor der Metamorphose bestanden sie grossenteils aus Schiefermergeln und Schiefertonen, sowie aus sandigen und mergeligen Kalken. Die Ablagerungstiefe der Radiolarite und der mit ihnen verknüpften Gesteine wird diskutiert, wobei die tiefmeerische Entstehung dieser Bildungen für wahrscheinlich angesehen wird. Marine Breccien sind bezeichnend für steile, oft durch Verwerfungen bedingte Geosynklinalränder, nicht aber für eine bestimmte Meerestiefe. Schliesslich wird eine Charakterisierung der alpinen Flyschbildungen versucht. Es sind dies mächtige, ausschliesslich marine, vorwiegend detritische Ablagerungen. z. T. Turbidite. Sie führen gewöhnlich keine vulkanischen Einschaltungen. Sie wurden während des beginnenden Zusammenschubes der Geosynklinale gebildet. Es gibt viele verschiedene Arten von Flysch, sowie Übergangstypen zu Bündnerschiefer und zu Molasse. Die wesentlichen Unterschiede zwischen Flysch und Molasse werden kurz aufgeführt.

Der zweite Teil der Arbeit gilt der Entwicklung der alpinen Geosynklinale. Argand's Theorie der Embryonalfaltung wird dargelegt, im wesentlichen aber abgelehnt. Die Trias entspricht einer "neutralen Periode" zwischen dem herzynischen und dem alpinen Zyklus. Während einer ersten Phase der Geosynklinalgeschichte herrschten Vertikalbewegungen, wahrscheinlich auch Dehnung, entlang normalen Verwerfungen vor. Diese Bewegungen wurden im Oberjura und in der untersten Kreide schwächer; es herrschte eine Zeitlang geringes Relief des Meeresbodens, bei "para-ozeanischen" Ablagerungsbedingungen in einem weiten Bereich des Mittelmeerraumes. Neue, langgestreckte Hochzonen, die ihre

Entstehung dem beginnenden Zusammenschub verdanken, entstanden während der Kreide; sie entwickelten sich zu den sogenannten "Kordilleren," steilen Inselrücken, welche die einzelnen Flyschbecken voneinander trennten. Der zentrale Teil der Alpen hob sich allmählich über den Meeresspiegel, die Geosynklinale wurde nach N und W abgedrängt und wurde schlussendlich zum Vorlandtrog des Molassebeckens.

Es besteht keine einfache, unmittelbare, räumlich-genetische Beziehung zwischen herzynischen Strukturen, Trögen und horstartigen Schwellen der ersten Geosynklinalphase, Becken und Kordilleren der "Flyschzeit," und der späteren Deckentektonik. Die palaeogeographischen Einheiten sind oft recht kurzlebig und verlagern sich im Lauf der geosynklinalen Vorgeschichte.

ПАЛЕОТЕКТОНИЧЕСКАЯ ЭВОЛЮЦИЯ ЦЕНТРАЛЬНЫХ И ЗАПАДНЫХ АЛЬП

Рудольф Трюмпи

Резюме

В статье рассматриваются общие черты разрезов мезозойских и третичных пород и палеогеография Альп. Статья ставит своей задачей описать в общих чертах значение горных пород в палеотектонике и обсудить структурную эволюцию Альпийской геосинклинали, включая главную деформацию, особенно в секторе между Рейном и рекой Дюранс. Особое внимание уделяется относительной самостоятельности альпийских структур, включающих дотретичный фундамент и покровы из мезозойских и третичных формаций.

Обычные мелководные осадки платформенного и миогеосинклинального типа были отложены на всей площади прежде чем наступили эвгеосинклинальные условия. Типичными эвгеосинклинальными отложениями в центральном, Пеннинском, поясе Альп являются блестящие кристаллические сланцы и кристаллические сланцы Бюнднер, с пластовыми интрузиями и подводными потоками основных вулканических пород (офиолитов). До метаморфизма эти отложения состояли, главным образом, из глинистых сланцев и нечистых песчанистых и глинистых известняков. Рассматриваются фациальные условия образования радиоляриевых кремнистых отложений с сопутствующими им породами. Подтверждается теория об их глубоководном происхождении в районе Альп. Наконец дается характеристика разрезов альпийского флиша. Флиш является мощным отложением преимущественно обломочных пород, частично отложением мутных потоков, обычно лишенным вулканических пород, и отложившимся во время компрессионной деформации геосинклинали. Из многих разновидностей флиша, некоторые представляют переходные типы к сланцу Бюнднер или к молассам. Перечисляются существенные различия между ранним синорогенным флишем и поздними синорогенными молассами.

При обсуждении вопроса об эволюции геосинклиналей рассмотрена и отвергнута теория Аргана о зародышевой тектонике. Триас соответствует нейтральному перерыву между герцинским и альпийским циклами. Раннеальпийская геосинклинальная история отличалась вертикальными движениями или движениями растяжения вдоль нормальных сбросов, которые окаймляли узкие платформы и быстро опускающиеся впадины. Эти деформации растяжения ослабели в течении поздне-юрского времени, вследствие чего разница в рельефе дна уменьшилась и параокеанические условия преобладали на значительной части средиземноморской области. В течении мелового периода возникли новые линейные поднятия компрессионного происхождения, превратившиеся в цепь крутых островов, обрамляющих флишевые бассейны. Постепенно центральная часть Альп поднялась над уровнем моря и геосинклиналь переместилась к северу и к западу. Здесь она была окончательно заполнена и уступила место предгорной впадине, в которой отлагались молассы.

Не существует прямого, генетического и пространственного сродства между герцинскими структурами, платформами, обрамленными сбросами, и впадинами ранних фаз развития геосинклинали, цепочками островов и бассейнами флишевой фазы, и позднейшими покровными структурами. Палеогеографические особенности обычно кратковременны и легко перемещаются.

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INTRODUCTION

Opening Remarks

There was a time when the Alps were considered, especially by European geologists, a model for all mountain chains. This was during the first quarter of the century, after the discovery of the Alpine nappe structure. For a while, nappes were found all over the world, even in places where they certainly did not exist. Ultranappism brought forth antinappism—which is only now losing some of its exaggerated violence—and the Alpine nappes were looked at as something quite exceptional. Their very existence was even questioned. At best, the Alps were a freak, not worth the trouble of trying to understand their history and structure.

The Alps, however, are neither the paradigm of all mountain chains nor a regrettable accident of the Earth's surface. They represent an extremely complicated but not a unique type of tectonic structure. Many tectonic phenomena can better be studied in different, less disturbed fold belts, but for others the outstanding known examples are in the Alps, with their deep valleys and their long history of investigation.

The classic era of Alpine geology lasted from about 1890 (Marcel Bertrand's discovery of the Glarus nappes, Schardt in the Prealps) to about 1925 (Staub's *Bau der Alpen*, 1924). Since then, a great amount of sometimes excellent local work has been done, but the day of the great synthetic views seems past, at least momentarily. During the last 15 years, the French geologists have made the most significant advances in the study of the Alps. The outlines of the structure are by now fairly well established, and detailed maps, mostly on scales between 1:25,000 and 1:75,000, cover almost the entire chain. Today, two problems are foremost: one concerns metamorphism and rock deformation, the other the sedimentary pre-history of the fold belt and its relationship with the later structures. This paper aims to indicate our state of knowledge on the second point.

Alpine geological writings of the last 2 decades, indeed, are not very encouraging to the extra-Alpine reader. The only general works in English (Collet, 1927; Bailey, 1935) are somewhat outdated, although the latter still makes delightful reading. The remarkable chapters devoted to the Alps in Gignoux's textbook[1] give a very clear picture for the French part of the chain, but progress in this part has been so rapid these last years that some major corrections of this picture are needed. In German, there are two excellent treatises, on the Swiss Alps by Cadisch (1953) and one on the Austrian Alps by Spengler and by Heritsch (*in* Schaffer and others, 1951); but both are essentially meant as reference books. For all details concerning structure and stratigraphic sequence the reader can be directed to these two standard works and to the great monograph by Albert Heim (1919; 1921), which contains a wealth of significant data.

Finally, inasmuch as they figure prominently in any discussion of Alpine geology, it seems appropriate to comment briefly here on some Alpine rock terms that have come to be used all over the world. Swiss geologists should be proud that their little country has given birth to terms like Flysch, Molasse, and Bündnerschiefer, which are recognized, or thought to be recognized, in almost any folded mountain chain. In fact, they are not as happy about it as they ought to be. These words have a perfectly sound household meaning in the Swiss Alps, but everybody is reluctant to give a categorical definition, and they were not meant for export. If "Flysch" and "Molasse" are to be used outside the Alps, these terms should at least designate rock groups showing some resemblance to the type Flysch and the type Molasse, and not merely as synonyms of pre-orogenic and post-orogenic deposits.

Acknowledgments and Limitations

This paper was written at the suggestion of Preston Cloud. For his unfailing patience and the pains he has taken in preparing the copy for print, the author is deeply indebted.

Marcel Lemoine, Walther Nabholz, and

[1] The English translation was not available to the author; references are to the last French edition of 1950.

John Rodgers have read the manuscript and have given invaluable help by their pertinent and constructive criticism. Augusto Gansser and my father, Daniel Trümpy, have also proffered very useful suggestions.

Two of my students have been very helpful in preparing the paper: Sibylle Dollfus for checking and compiling references, Kimon Karagounis for drawing most of the figures.

A paper of this kind is necessarily based on the work of a great number of contemporaries and predecessors in the field reviewed. Because much of this, however, is so well known as to amount to common knowledge in the case of the Alps, it is unnecessary as well as impracticable to cite specifically all works which contain relevant contributions to Alpine geology. It would also be unrealistic to enter into long discussions on all points of disagreement. The author is painfully aware that this paper contains some rather unequivocal statements on questions which are in fact still disputed; but a certain amount of this is unavoidable when a large field is compressed into small scope, and one can only hope that time will vindicate his choice of subjects to slight.

A major shortcoming of the present paper is its unequal treatment of the different parts of the Alps. Most examples are taken from the northern miogeosyncline (Helvetic-Dauphiné realm) and from the central, Penninic, eugosyncline. The author is less familiar with the highest, Austroalpine, nappes of Austria and Germany and fears to misjudge the significance of local facts, in spite of all the important papers written on this part of the chain. This, again, is a risk which is inherent in the preparation of such a summary paper, and the author can only ask forebearance on the part of colleagues who may know better.

Outlines of Tectonic Structure

Nappe structure and the amount of crustal shortening.—The nappe structure of the Alps can no longer be questioned, but the mechanism, the correlation, and the amplitude of the nappes are still controversial. Uncertainties in correlation fundamentally affect our views on paleogeography and render the construction of palinspastic maps and sections through the whole fold belt difficult and problematical. A French geologist with a tendency to ignore the existence of the Valais facies belt and to claim Piemont origin for all Schistes lustrés, a Swiss geologist who regards most Prealpine nappes as lower Austroalpine units, and an Austrian geologist who maintains a local origin for the northern calcareous Alps east of the Rhine would develop different views on the early history of the Alps from the ones sketched here. It is inevitable that one should take a personal stand on these questions, from the best of one's knowledge, even in a would-be dispassionate review article.

Widely differing estimates have been reached for the amount of crustal shortening in the Alps. Cadisch (1953, p. 287) gives a pre-folding breadth of 630 km for the Swiss Alps, which now measure only 150 km across. This figure, obtained by straightening out one folded stratigraphic datum along a cross section, is open to criticism on two points:

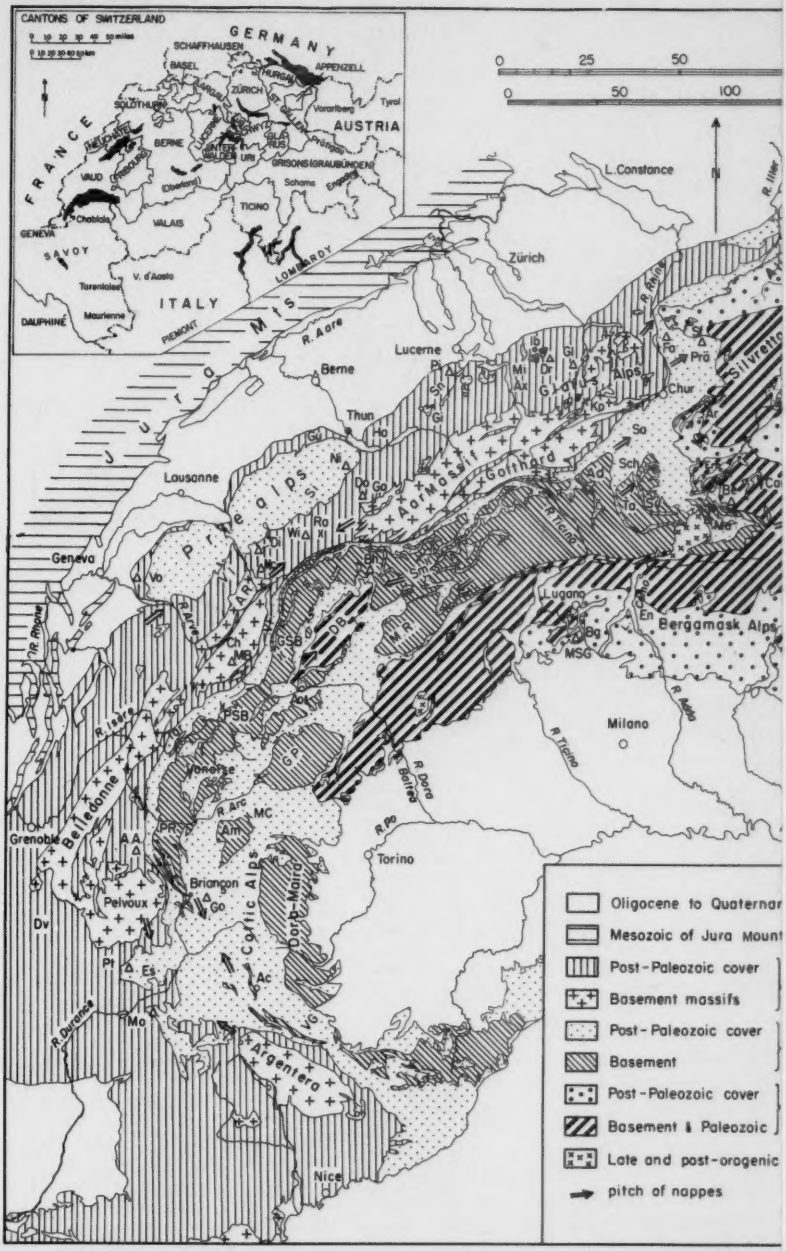
(1) It neglects the stretching of Alpine rocks, which may attain proportions of five to one or more in certain zones, especially in the deeper-seated parts of the chain. This stretching is probably not quite compensated by microfolding.

(2) Structural and paleogeographic analysis shows that several important units have disappeared. Many have been removed by erosion; the volume of the detrital Oligocene-Miocene Molasse alone is as great as that of the present Alps. Other structural elements have been drawn downward, into the inaccessible (and presumably granitized) depths of the substratum.

The two sources of error work in opposite directions, so that Cadisch's figure of 630 km for the initial breadth of the Alps may be taken as a fair approximation (±200 km). A 50 per cent shortening, from 300 to 150 km, probably represents an absolute minimum. (*See* also Sonder, 1940.)

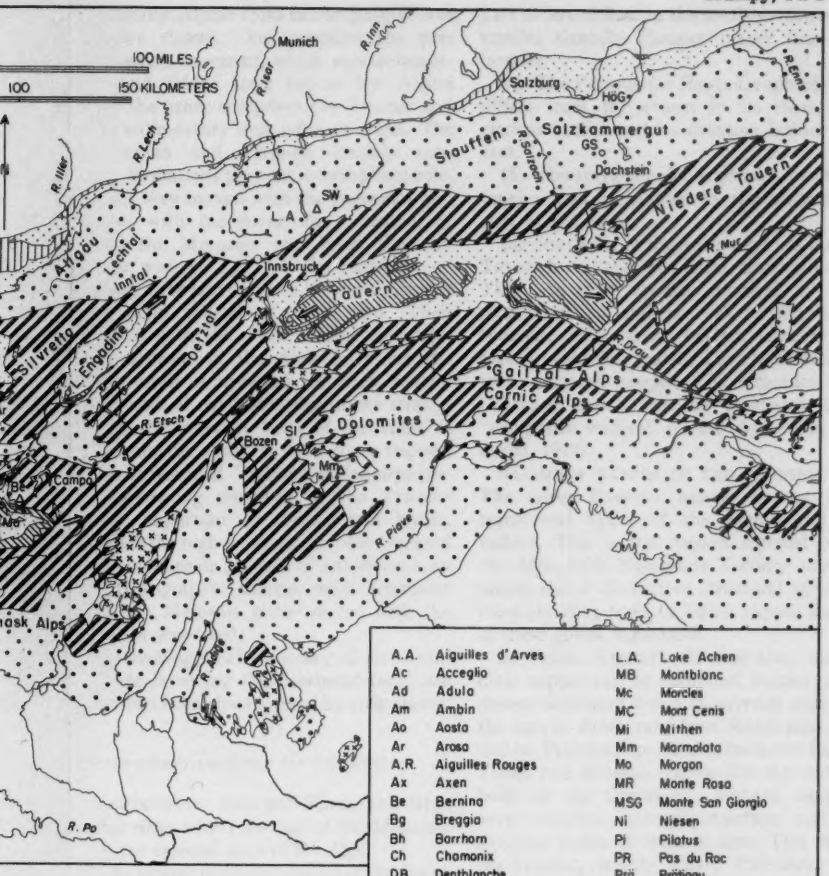
Nappes are generally ascribed to overthrusting (rather than overfolding) from the south, or from the east in the French part of the chain. Underthrusting from the north, however, seems to provide a more satisfactory cause for many phenomena, although most structures can be explained in either way. In discussing this and other matters it is convenient to refer to units belonging originally to a more southerly or easterly belt as internal and, conversely, to consider the northerly or westerly zones as external.

For the structural pattern of the Alps, the reader is directed to the standard works mentioned above. Only a few generalities are given here as the minimum essential background for the discussions to follow. Plate 1 shows the broad areal distribution of the principal tectonic units.



TECTONIC SKETCH

TECTONIC SKETCH MA



A.A.	Aiguilles d'Arves	L.A.	Lake Achen
Ac	Acceglio	MB	Montblanc
Ad	Adula	Mc	Morcles
Am	Ambin	MC	Mont Cenis
Ao	Aosta	Mi	Mithen
Ar	Arosa	Mm	Marmolata
A.R.	Aiguilles Rouges	Mo	Morgon
Ax	Axen	MR	Monte Rosa
Be	Bernina	MSG	Monte San Giorgio
Bg	Breggia	Ni	Niesen
Bh	Barrhorn	Pi	Pilatus
Ch	Chamonix	PR	Pas du Roc
DB	Dentblanche	Prä	Prättigau
Di	Diablerets	PSB	Petit St. Bernard
Do	Doldenhorn	Pt	Piolit
Dr	Drusberg	Ra	Rawilpass
Dv	Devoluy	Sa	Safien
En	Esino	Sch	Schams
Es	Escoureous	Sf	Sulzfluh
Fa	Falknis	Si	Simme
Go	Gastern	Sl	Schlern
Gi	Giswiler Stöcke	Sm	Simplon
Gl	Glärnisch	Sn	Schlieren
Go	Mont Gondran	Su	Suretta
GP	Gran Paradiso	Sw	Sonnwend Mts.
GS	Gosau	Ta	Tambo
GSB	Grand St. Bernard	Tar	Tarentaise
Gu	Gurnigel	VF	Val Ferret
Ha	Habkern	VG	Val Grana
HöG	Höllengebirge	VI	Val d'Illiez
Ib	Iberg	Vo	Voirons
Kp	Kistenpass	Wi	Wildhorn

Quaternary cover

ura Mountains

cover
ssifs } Helvetic & Ultrahelvetic

c cover } Penninic belt

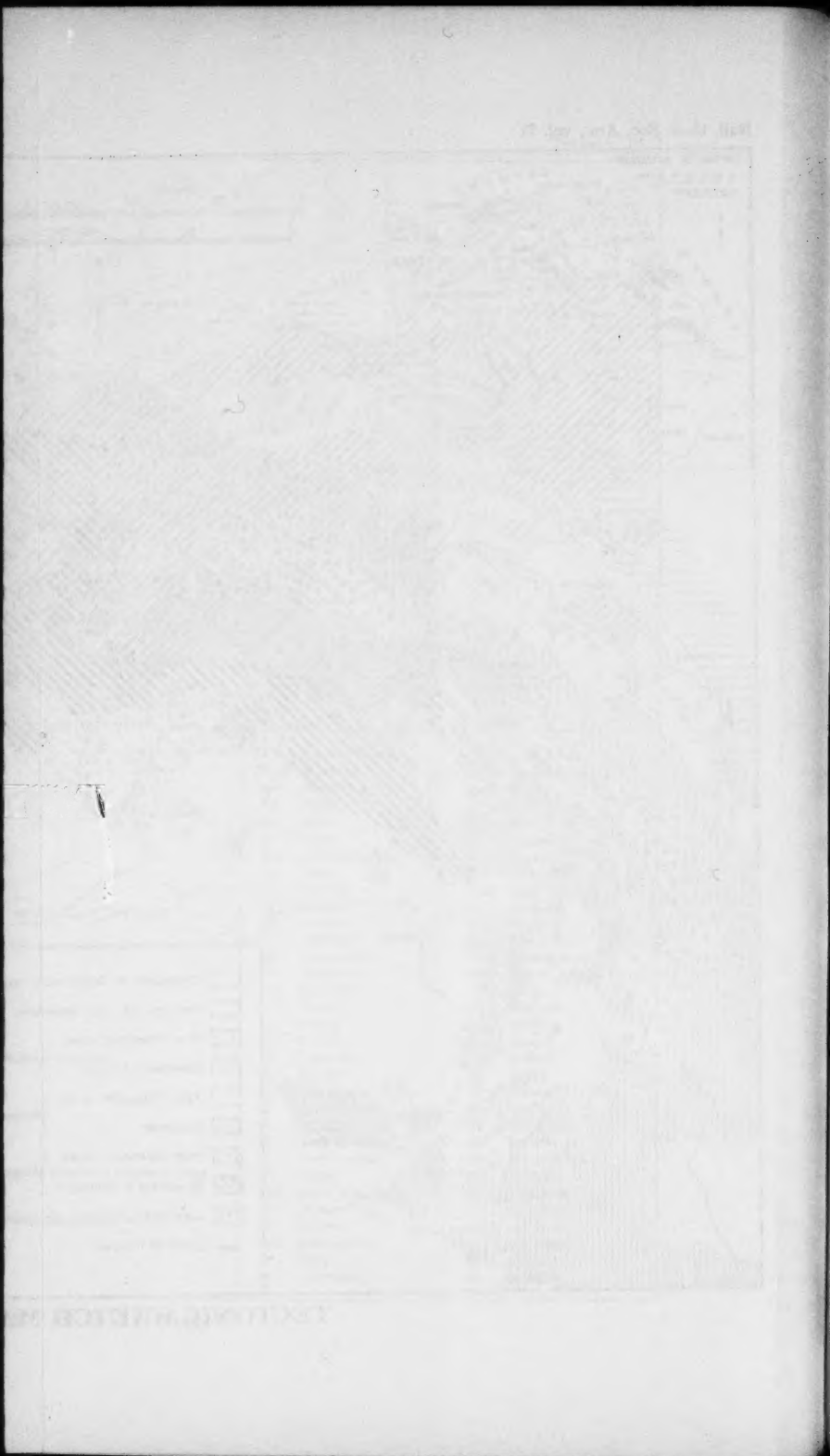
cover
aleozoic } Austroalpine nappes & Southern Alps

-orogenic granites & volcanics

es

TCH MAP OF THE ALPS

1960
UMI



Structurally, Alpine rocks can be grouped into two major classes. One contains the pre-Pennsylvanian basement, which was metamorphosed and folded long before the Alpine orogeny; the other comprises the Triassic and younger sedimentary and volcanic rocks. The Pennsylvanian and Permian (mainly continental) formations play an intermediate role; sometimes they moved with the basement and sometimes with the cover. In a number of instances the Mesozoic and lower Tertiary sediments were stripped off the basement on which they were laid down to form décollement nappes (strip sheets) of their own. Their relationship with the much thicker nappes of the central zone of the Alps, in which the basement itself is involved, is often controversial. Much of our knowledge of the stratigraphy of the internal belts of the chain is derived from the study of such cover nappes, which have escaped Tertiary metamorphism. Anhydrite-bearing members in the Triassic[3] are most important as incompetent layers, along which the higher formations were sheared off; and presence or absence of anhydrite beds in the stratigraphic column may determine whether the Mesozoic cover stayed with the basement or wandered.

In the ensuing brief summary of structural units those involving the basement rocks are distinguished from those formed by strip sheets.

Alpine structures involving the basement.—

AUTOCHTHONOUS MASSIFS: These are large basement slices which rise out of the Mesozoic cover in the external zone of the Alps.

(1) The largest is the Aar massif of the Berne and Uri Alps, in central Switzerland. The small, cylindrical Gastern massif to the north is only locally separated from the main body of the Aar massif by a narrow wedge of Mesozoic sediments.

(2) West of a transversal depression, Alpine equivalents of the Aar massif rise again in the Montblanc massif. The Aiguilles Rouges massif, beyond the Chamonix "syncline" to the northwest, occupies the structural site of the Gastern massif.

(3) The long and bifid Belledonne massif prolongs both Aiguilles Rouges and the external part of Montblanc to the south-southwest. The smaller Grandes Rousses massif lies to the interior.

(4) The Pelvoux (or Haute Dauphiné) massif differs from the others in its more nearly circular outline and its situation farther to the east.

(5) The amygdaloidal Argentera (or Mercantour) massif is isolated in the southernmost part of the arch.

(6) Southeast of the Aar massif lies the small Tavetsch and then the larger Gotthard massif. They are more strongly marked by Alpine metamorphism and are not strictly autochthonous; they have moved laterally with regard to the Aar massif. They have even been interpreted as nappes. The structure of the Gotthard massif can be compared to the "composite wedges" of the Apennines (Migliorini, 1948).

BASEMENT NAPPES OF THE PENNINIC BELT: The lower Penninic nappes appear in the transversal uplift of the Ticino and Toce valleys. This is the deepest exposed part of the Alps, with high-grade Tertiary metamorphism and a distinctive structural style. The steep-standing roots of higher nappes lie south of these gneiss mountains.

In Valais, Argand's classical area, the Penninic nappes can be examined, thanks to their descent westward into a transversal depression, the axis of which runs from Rawil pass to the central Pennine Alps. Above the lower Penninic Ticino and Simplon nappes lies the enormous body of the Grand St. Bernard nappe, of very complex internal structure and very irregular strike of tectonic axes. This mass of pre-Triassic, largely upper Paleozoic, rocks continues down to the Mediterranean, but because its base is not exposed many geologists doubt its nappe character in the western Alps. A broad belt of cover nappes, however, has been thrust out between the autochthonous massifs and the Grand St. Bernard, so that these two basement units must have been laterally displaced toward each other.

To the inside and apparently above the Grand St. Bernard lies the Monte Rosa nappe, with a plunging front. The base of the similar gneiss cupolas of the Gran Paradiso and of the Dora-Maira is not exposed.

In Graubünden, pitch toward the east again shows a natural cross section through all the Penninic nappes. Correlation of units on the two sides of the Ticino uplift is often difficult, and most geologists now working in the Ticino

[3] Represented at the surface by cornieules (cargneules, Rauhwacken), very porous and brecciated rocks consisting mainly of calcite, instead of the original anhydrite dolomites. Regarding the genesis of cornieules, *see* Brückner (1941).

tend to discard the possibility. There is, however, a kind of broad correspondence between basement nappes of the Valais and Graubünden. The Adula gneiss nappe is the main unit of the lower Penninic. Two thick, superposed nappes, Tambo and Suretta, Staub correlates with the Grand St. Bernard and Monte Rosa nappes of Valais, although the structures are very different on the eastern and western sides of the Ticino uplift. The Margna nappe (Ma on Pl. 1), severely sliced up under the Austroalpine thrust plane, is the most important basement nappe of the upper Penninic units.

The Penninic nappes then disappear to the east under the great Austroalpine thrust mass. Their Mesozoic cover is again exposed in the window of the Lower Engadine, and both cover and basement reappear in the window of the Hohe Tauern, in the center of the eastern Alps. To any geologist acquainted with the whole chain, the Penninic nature of the elements inside the Tauern window is beyond doubt, even if the Penninic zone may have "gone autochthonous" and lost some of its nappe structure this far east. Clar (1953) gives an excellent short review on the importance of the Tauern window for the structure of the eastern Alps.

BASEMENT NAPPES OF THE AUSTROALPINE BELT: The lower Austroalpine basement units of Graubünden, which contain much Hercynian granite and granodiorite, are represented mainly by the Err and Bernina nappes. In the Valais Alps, the large thrust mass of the Dentblanche, which rests everywhere on Penninic Mesozoic rocks, can very probably be reckoned as an approximate equivalent. It is not preserved southwest of the Aosta Valley but must once have overlain most of the western Alps.

The "Central Alps" of Austrian geologists, east of the Rhine, are formed of crystalline basement rocks, upon which Paleozoic formations are preserved in large tracts, especially on the northern, eastern, and southeastern border. They constitute the basement core of the upper Austroalpine nappes (including the ill-defined "middle Austroalpine" of Graubünden), partly split up into the Campo, Silvretta, and Oetztal nappes; the separations between these units cannot be followed over a long distance. Wedges of Triassic rocks are also found in Tyrol and Styria. At their western termination, and along the "frame" of the Tauern window, the upper Austroalpine basement rocks rest upon the Mesozoic to lower Eocene sediments of Penninic and lower Austroalpine units; but to the south, the limit of the certainly autochthonous basement of the southern Alps can be drawn only with difficulty. This seemingly paradoxical relationship of the greatest thrust sheet of the Alps with an autochthonous belt is more easily understood if one takes underthrusting from the north as the main movement of the Alpine orogeny. The upper Austroalpine basement nappes did not extend much farther west than their present erosion edge in central Graubünden.

Décollement nappes.—

HELVETIC NAPPES AND SUBALPINE CHAINS OF FRANCE: In the Swiss Alps, the cover of the Aiguilles Rouges and Gastern massifs is strictly autochthonous, except for some independent sliding of the Flysch group. The Mesozoic and Eocene cover of the Montblanc and of the western Aar massif is stripped off and forms a large recumbent fold, the Morcles or Doldenhorn nappe. In eastern Switzerland, where there are no anhydrite members in the Triassic, the cover of the Aar massif forms only parautochthonous slices, little displaced with respect to their substratum. The Helvetic nappes proper are derived, in the east, from the former covering beds of the Tavetsch massif and, perhaps, from the northern part of the larger Gotthard massif (Fig. 1A; Pl. 1). They are formed by a number of very spectacular although rather small nappes (*e.g.*, Axen and Drusberg on Lake Lucerne, Diablerets and Wildhorn in western Switzerland); but their correlation is hazardous, as the location of thrust planes is controlled largely by sedimentary facies. The Ultrahelvetic nappes, probably derived from atop the Gotthard massif and its buried western equivalents, slid onto the back of the future Helvetic nappes in Oligocene times. When the Helvetic nappes were formed, in the Miocene, the Ultrahelvetic nappes became pinched into the recumbent synclines between and below the Helvetic elements, and their internal structure is now hopelessly chaotic.

On the river Rhine the Helvetic nappes and the Aar massif pitch very steeply east, below Penninic and Austroalpine units; but a narrow strip of Helvetic and Ultrahelvetic elements continues along the northern border of the eastern Alps, below the Penninic Flysch zone.

In the French Alps of Savoy and Dauphiné, the Helvetic nappes have lost their individuality. Southeast of lake Geneva, the Morcles

fold also merges with the Subalpine chains, of which the main body is a continuation of the autochthonous zone of Switzerland extending finally under the Molasse basin. These Subalpine chains form a broad, folded belt of Appalachian style. The Mesozoic formations are probably sheared off the basement along the lubricating anhydrite layers of the Triassic, and the outermost of these folds join the southern end of the Jura Mountains. In southeastern France, Eocene ("Pyrenean") east-west folds complexly interfere with the Alpine folds of Miocene age, and the Triassic anhydrite rises in diapiric plugs.

The Ultradauphinois slices, behind the Belledonne and Pelvoux massifs, are more or less in the prolongation of the Ultrahelvetic root zone of the Valais Rhone valley.

PENNINIC COVER NAPPES OF THE FRENCH ALPS: In the depression between the Pelvoux and Argentera massifs, Penninic décollement nappes rest upon the Jurassic shales of the Subalpine Dauphiné belt. This is the type area of the Subbriançonnais, with the two main digitations of Piolit and Morgon. Above is a strip sheet showing Briançonnais facies (Escouréous), and this in turn is overlain by the great mass of the upper Cretaceous Helminthoid Flysch, also present in the Maritime Alps to the southeast, and of still more internal (Piemontese or Austroalpine?) origin.

Farther east, the Mesozoic of the type Briançonnais, in the upper Durance Valley, is less markedly separated from its original substratum, the Pennsylvanian formations of the Zone houillère along the front on the Grand St. Bernard mass. Nevertheless, it forms several small nappes and is overridden by the Schistes lustrés from the Piemont realm. All through the western Alps, these Schistes lustrés have moved as one or several cover nappes, independently of their former basement, which may have been the Dora-Maira and Gran Paradiso gneiss domes or even more internal elements.

To the north, in the Arc and Isère valleys of Savoy, the Subbriançonnais units are mainly represented by the Pas du Roc nappe, and other nappes not visible in the Durance valley lie west of it. The Mesozoic cover is gone from the Zone houillère, probably stripped off to form part of the Prealps; but it is preserved on a more internal digitation of the Grand St. Bernard mass, in the Vanoise Mountains and around Mont Cenis.

PREALPINE NAPPES: The Prealps consist of two semicircular areas along the front of the Alps, between the river Arve and Lake Thun. Erosion has spared other outliers of Prealpine nappes to the south, in the klippen of Savoy, and to the northeast around Lake Lucerne. A cross section through the Prealps is given in Plate 2.

From the lowest to the highest, the following units are represented in the Prealpine pile of cover nappes:

(1) Ultrahelvetic nappes, which crop out along the inner margin northwest of the Helvetic high calcareous Alps (Préalpes internes) and along the border of the Molasse basin (Préalpes externes)

(2) Isolated slices, like the Chamossaire nappe, and (Penninic?) Schlieren Flysch nappe of central Switzerland

(3) Niesen nappe, of unknown, presumably lower Penninic, origin

(4) The nappe of the Préalpes Médianes or Klippen nappe, derived from the cover of the middle Penninic Grand St. Bernard mass.[3] The northwestern part (Médianes plastiques) has Subbriançonnais facies; the more internal Médianes rigides belong to the Briançonnais proper

(5) Breccia nappe, which is discussed under Marine Breccia Formations

(6) An ophiolite-bearing unit, preserved only locally, may be correlated with the high Penninic Platta nappe and the Arosa zone of Graubünden

(7) Simme nappe (Austroalpine?), consisting mainly of Cretaceous Flysch. This unit formerly had a great extent and furnished most pebbles to the Oligocene conglomerates of the Molasse

(8) Erosion remnants of an upper Austroalpine nappe exist northeast of Lake Lucerne

The order of movement can be described as follows:

(1) Sliding of Simme and Ophiolite nappes onto future Breccia and Médianes nappes; sliding of Ultrahelvetic nappes (at the same time?) onto future Helvetic nappes and onto the Autochthone. Gravity seems to have been an essential factor in this early (Oligocene) phase

(2) Main thrusting of the Médianes nappe

(3) Arrival of the Breccia nappe on top of the Médianes, and (at the same time?) formation of the Helvetic nappes. The Morcles fold developed before the Helvetic nappes proper.

[3] Thanks to Ellenberger's (1949; 1958) discoveries of the typical Médianes rigides formations in the Vanoise, the long-disputed question of the origin of this unit is now probably settled, in spite of Staub's (1958) impressive defense of its lower Austroalpine character.

(4) Passage of the Niesen nappe through the transversal depression between the Montblanc and Aar massif

(5) Late en-bloc deformation of the Prealps

COVER NAPPES OF GRAUBÜNDEN: In Graubünden (Grisons), there is a great mass of lower Penninic Bündnerschiefer and Flysch, which reappears in the windows of the Lower Engadine and of the Tauern, as well as in the Flysch belt along the northern border of the Bavarian and Austrian Alps. These Mesozoic to Eocene, partly metamorphosed formations are overlain by other cover nappes—those of the Schams in the south and the three units of Falknis, Sulzfluh, and Arosa in the north and east. The Sulzfluh nappe is probably the Médianes nappe of the Prealps. The author regards the ophiolite-bearing Arosa zone as identical with the high Penninic Platta nappe, and all three nappes as Penninic, but this view is not shared by many geologists who still attach these units to the lower Austroalpine nappes. In the present interpretation the Mesozoic cover of the Err and Bernina nappes has not moved far from its pre-Triassic basement in the Upper Engadine.

AUSTROALPINE COVER NAPPES OF THE NORTHERN CALCAREOUS ALPS: These nappes form a broad, continuous belt between the Rhine and the Danube. On their southern border, they lie upon or butt against the great mass of pre-Triassic rocks of the upper Austroalpine thrust sheet, from the cover of which they have probably been stripped along incompetent layers in the Paleozoic and Lower Triassic formations. The sedimentary facies of all these nappes is very similar (except the Hallstatt nappe), and the distinction and correlation of the individual sheets are hazardous. In the west, however, there are three well-defined units: Allgäu, Lechtal, and Inntal. Lechtal is the most important and more or less continues eastward into the Stauffen-Hollengebirge nappe of the Salzburg Mountains. East of the river Enns, beyond the border of Plate 1, three separate thrust sheets replace it.

In the central segment of the northern calcareous Alps, especially in the Salzkammergut, the highest cover nappes are preserved. The most voluminous of these is the Dachstein ("high-Alpine") nappe. Two units not shown on Plate 1, the Hallstatte nappe (reduced to isolated slices), remarkable for its aberrant facies of the Triassic, and the overlying Reiteralm nappe, have moved during the Cretaceous pre-Gosau phase, but the main disturbance is Tertiary. Whether the Hallstatt nappe is of more southerly origin than the Dachstein—which would result in the most plausible picture of facies distribution—or whether it is more northerly is still being disputed. Fallot (1954) gives the clearest account of the structural problems of the eastern Alps.

STRUCTURE OF THE COVER IN THE SOUTHERN ALPS: The structural style of this area is much simpler than that of the other Alpine regions, and the apparent sense of movement is reversed (slight overthrusting from the north or underthrusting from the south, along steeply inclined planes). Oblique faults, in part slip faults, are prominent. Especially the youngest, post-Triassic formations are locally affected by gravity sliding southward.

Outlines of Stratigraphy

Facies belts.—In the Alps, we can distinguish five major facies belts, each with a different geological history (Fig. 1A; Pl. 1). They are, from the northwest to the southeast:

(1) Helvetic realm of Switzerland and Dauphiné realm of southeastern France: external miogeosyncline. Its appendage, the Ultrahelvetic and Ultradauphinois belt, leads over to

(2) Valais facies belt: external eugeosyncline. It is relatively narrow in the Valais itself. In the east (Graubünden) this facies realm becomes much broader and more complex, whereas it may have disappeared to the southwest, in the Durance basin.

(3) Briançonnais belt in the large sense: subdivided into the external Subbriançonnais and the internal Briançonnais in the strict sense.

(4) Piemont belt of northwestern Italy: internal eugeosyncline, narrowing toward the northeast. This and belts (2) and (3) constitute the Penninic realm.

(5) Austroalpine and south Alpine realms: relatively stable, but at times deeply submerged area beyond the Penninic eugeosyncline. The transitional zone between this and the Piemont belt is formed by the Grisonide (Graubünden) belt of the lower Austroalpine nappes.

The stratigraphic composition of these belts is outlined for the benefit of the extra-Alpine reader and with apologies to the Alpine reader for the gross simplifications. As a convenience to those who might otherwise have to look them up the European stage names for the Mesozoic are given in Table 1.

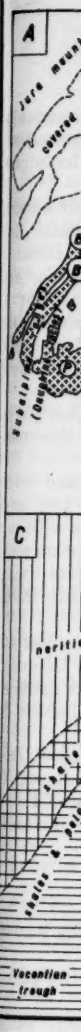


FIGURE 1.—P

(*See*

A. Original
pattern of p
AM Argenter
P Pelvoux
external

Only B^1, Al
for the Tertiar
nappes is also
B. Early Ju
C. Early C
D. Late Cre

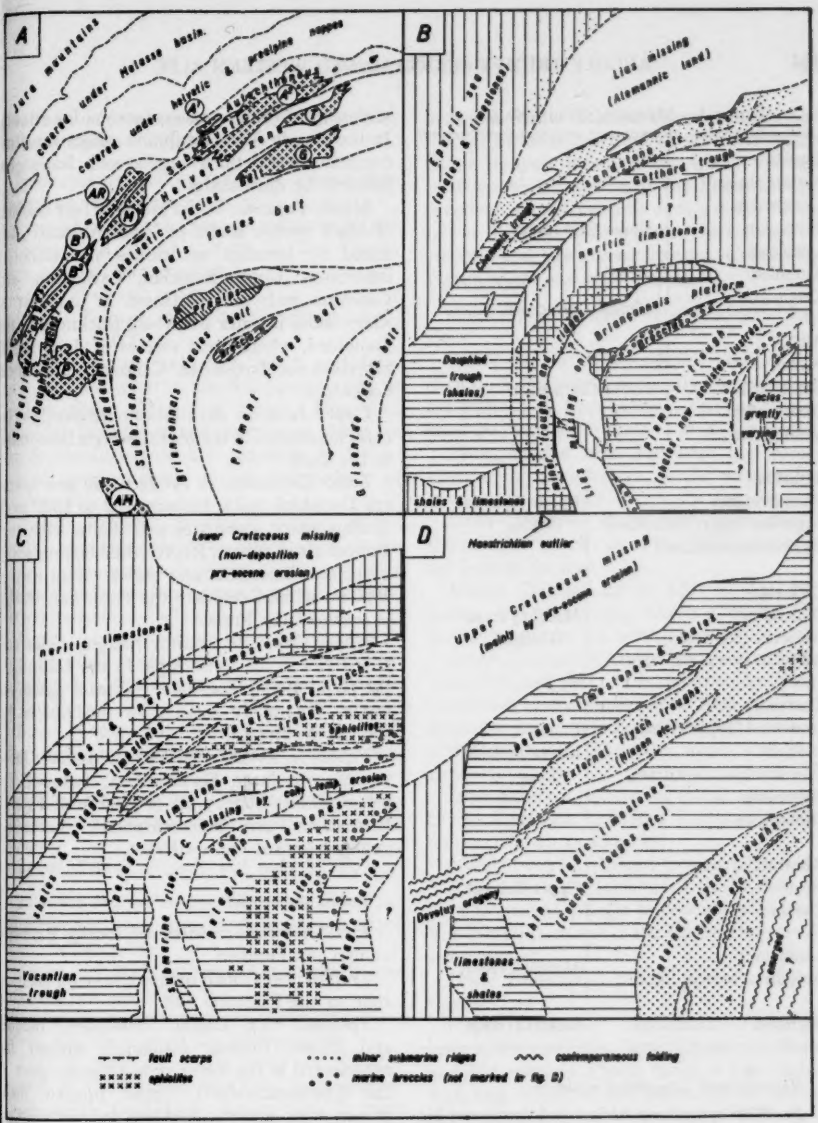


FIGURE 1.—PALINSPASTIC SKETCH MAP (A) AND PALEOGEOGRAPHICAL SKETCHES (B-D) OF PART OF THE WESTERN ALPS

(*See* Plate 1 for orientation). From R. Trümpy (1958, p. 342) and very hypothetical

A. Original disposition of facies belts. The more or less autochthonous basement massifs are marked by a pattern of plusses:

AM	Argentera or Mercantour	B^2 internal branch of Belledonne	A^1 Gastern and
P	Pelvoux		A^2 main part of Aar
E	external and	AR Aiguilles Rouges	T Tavetsch
		M Montblanc	G Gotthard

Only B^1, AR, and A^1 are shown in their present position; the other units have been displaced to account for the Tertiary deformations. The probable area of deposition of the sediments of the Médianes and Breccia nappes is also indicated.

B. Early Jurassic (Lias) paleogeography

C. Early Cretaceous paleogeography

D. Late Cretaceous paleogeography

TABLE 1.—MESOZOIC STAGE NAMES

Danian		
Maestrichtian		
Campanian		
Santonian	Upper Cretaceous	
Coniacian		
Turonian		
Cenomanian	Middle Cretaceous	
Albian		
Aptian		
Barremian	Lower Cretaceous	
Hauterivian		
Valanginian		
Tithonian		
Kimmeridgian	Upper Jurassic	
Argovian (upper Oxfordian)	(Malm)	
Oxfordian (restricted)		
Callovian		
Bathonian	Middle Jurassic	
Bajocian (restricted)	(Dogger)	
Aalenian (lower Bajocian)		
Toarcian		
Domerian (upper Pliensbachian)	Lower Jurassic	
Pliensbachian (restricted)	(Lias)	
Sinemurian		
Hettangian		
Rhaetian		
Norian	Upper Triassic	
Carnian		
Ladinian	Middle Triassic	
Anisian		
Scythian	Lower Triassic	

Helvetic and Dauphiné realm.—

Pre-Pennsylvanian: folded and metamorphic basement group; granites

Pennsylvanian: up to 1400 m of continental, coal-bearing beds in narrow late Hercynian synclines; granites

Permian: in localized troughs, up to 1800 m of red beds with volcanic rocks

Triassic: thin, comprising quartzites, dolomites, and variegated shales. Includes anhydrite beds in the south and west; spilites at the top in part of the Dauphiné belt

Lias: missing in the north; in the Helvetic nappes it is represented by up to 500 m of shales, sandstones, and arenaceous limestones. In the French Alps (Dauphiné miogeosyncline) comprises very thick argillaceous limestones followed by dark shales

Middle Jurassic: in the Helvetic belt consists of black pyritic shales of lower Aalenian, followed by ironshot sandstones and crinoidal limestones. Upper Bajocian, Bathonian, and Callovian beds are condensed in the east. A shaly facies invades the whole thickness in the Dauphiné, where great masses of dark shales represent the Bathonian, Callovian, and lower Oxfordian.

Upper Jurassic: dark limestones overlain by coral limestones in the north, pelagic limestones to the south.

Lower Cretaceous: in the Helvetic and northern Dauphiné realm includes 100 to 1500 m of shallow-water limestones and shales with thin glauconitic members. Represented farther south by pelagic limestones and shales. Gives way to 2000 m of marl shales in the Vocontian trough of southeastern France.

Upper Cretaceous: mainly pelagic shales and limestones in the Swiss Alps. In the Dauphiné, the facies of the limestones indicates shallower water. Commonly missing or incomplete by pre-Eocene erosion.

Paleocene, Eocene, and lower Oligocene: limestones with large Foraminifera, followed by pelagic shales and by up to 1500 m of Flysch. Formations are heterochronous, becoming younger toward the northwest.

Valais facies belt.—

Pre-Pennsylvanian basement: locally Pennsylvanian and Permian

Triassic: very much reduced, save for passage zone to the following belt

Jurassic: thin Liassic limestones. Middle and Upper Jurassic commonly absent (or represented in the lower unfossliferous part of the Bündnerschiefer?). Upper Jurassic limestones with granite boulders in the Falknis nappe (southeastern margin of the belt)

Lower Cretaceous: main mass of north-Penninic Bündnerschiefer (2000–3000 m?), with ophiolitic volcanic rocks in the lower part. In the upper part, Flysch characteristics appear gradually ("pre-Flysch"). Typical members are Aptian limestones with dolomite debris (Tristel and Aroley limestones, often unconformable on older rocks), and the green quartzites and black shales of the so-called "Gault."

Upper Cretaceous, Paleocene, and Eocene:

Flysch several kilometers thick, belonging to different basins

Briançonnais belt in the large sense.—Pre-Pennsylvanian basement

Pennsylvanian and Permian: continental formations with volcanic rocks in the French Alps, several kilometers thick

Lower Triassic: quartzites

Middle Triassic: algal limestones and dolomites, up to 1000 m.

Upper Triassic: anhydrite, shale, and dolomite; missing in part of the Briançonnais proper

Lias: about 500 m of cherty, silty, and crinoidal limestones and calcareous shales in the Subbriançonnais belt. Missing in Briançonnais proper, except in the eastern part of the Vanoise.

Middle Jurassic: Zoophycos Dogger of the Subbriançonnais belt, thick formation of shales and silty limestones, in places oölitic and coral limestones. In the Briançonnais proper, *Mytilus* beds (Bathonian-Callovian), with coal

Upper Jurassic: mainly neritic limestones in the Prealps, transgressive upon granite in Graubünden. Pelagic limestones predominate in the French Alps.

Lower Cretaceous: pelagic limestones with chert bands, absent from most of Briançonnais proper

Upper Cretaceous and Paleocene: marly limestones with pelagic Foraminifera (Couches Rouges)

Lower Eocene: Flysch, but only locally

Piemont belt.—Pennsylvanian and Permian: found locally, but commonly difficult to separate from the basement

Lower and Middle Triassic: as in the Briançonnais, but much thinner

Upper Triassic: variegated shales and dolomites, thicker than in the Briançonnais

Lias to Bajocian: main body of Schistes lustrés in the western Alps (but note reservations as to their age). Lower Lias in part limestones, with benthonic fauna. Microbreccias along the margins of the trough. Thickness around 2000 m?

Bathonian? to Lower Cretaceous?: thin radiolarian cherts and fine-grained limestones. Locally Lower Cretaceous schists and breccias. Intrusion and outflow of basic and ultrabasic ophiolites during the Early Cretaceous.

Upper Cretaceous and Paleocene: Couches rouges facies on the external margin, Flysch farther inward.

The stratigraphic succession of the Breccia nappe (external margin of Piemont belt) is discussed under Marine Breccia Formations.

The *Grisonide facies belt* of the lower Austroalpine nappes shows on the whole the same stratigraphic succession as the Piemont eugeosyncline, but no volcanic rocks, much thinner Lower Jurassic, and a great development of Jurassic and Cretaceous breccias. Cenomanian Flysch is the youngest unit.

Austroalpine realm.—Pre-Ordovician basement

Ordovician to Mississippian: fairly thick marine formations, especially developed in the east (Graz, Carnic Alps). The age of many slate-graywacke formations is still uncertain.

Pennsylvanian and Permian: not present everywhere; continental in the western and central zone, partly marine in the southeast

Lower Triassic: shales and sandstones in the east (with rock salt of the Salzkammergut), red beds in the west

Middle Triassic: up to 1200 m algal limestones and dolomites; siliceous shales with volcanic tuffs in the west

Upper Triassic: Carnian shales, sandstones, dolomites, and plant beds, followed by Norian limestones or dolomites (up to 1000 m). In the Hallstatt nappe, much thinner pink limestones with ammonites. Rhaetian shales and limestones.

Lias: facies widely varying (calcareous shales, cherty limestone, crinoidal limestone, pelagic limestone with ammonites)

Middle Jurassic: shales, limestones, chert beds

Upper Jurassic: radiolarian cherts, followed by fine-grained pelagic limestones with chert. The overall thickness of the Jurassic is generally quite modest.

Lower Cretaceous: pelagic shales and limestones

Upper Cretaceous: Cenomanian Flysch-like beds in the west. The Upper Cretaceous Simme and Helminthoid Flysch facies of the western Alps may belong to this realm. In the east are the Senonian Gosau beds, up to 2000 m of sandstones, conglomerates, shales, and shallow-water limestones, locally with workable coal deposits, unconformable on older, folded rocks

Eocene and lower Oligocene: locally unconformable detrital formations

Realm of the southern Alps.—In general similar to that of the northern calcareous Alps. The Permian is mainly marine in the east and continental in the west; it contains much volcanic rock and also some granites (Baveno). Besides dolomites and algal limestones, the Middle Triassic also shows volcanic rocks and

euxenic limestones. The composite thickness of the Triassic may reach 3500 m. The Upper Cretaceous consists of pelagic marl shales, followed by Flysch.

Genetic Sequences of Alpine Sedimentary Rocks

Platform and Miogeosynclinal Sediments

Gross characteristics.—In the Alps, large areas of shelf and miogeosynclinal deposits have been affected by the nappe-building processes, and some of them by metamorphism of epizonal to mesozonal grade. (*See* especially Ellenberger, 1958, p. 353–405.) In these miogeosynclinal belts, the rate of subsidence and of sedimentation may be greater than in the eugeosynclines,[4] but the sediments as a rule are of normal, shallow-water origin. Changes in facies and thickness are commonly rather abrupt as compared to conditions in the "stable-shelf" areas outside the Alps, such as the Jura Mountains or Swabia.

The space–time relationship of miogeosynclinal and eugeosynclinal sequences in the Alps can be summarized as follows:

(1) Up to the Triassic or to the early Lias, miogeosynclinal or platform conditions prevailed over the whole area of the future Alps. The Triassic miogeosynclines, in the Briançonnais belt of the western Alps and in the Austroalpine to south Alpine realm, are in part inherited from late Paleozoic post-orogenic basins ("epieugeosynclines"). In any given section, eugeosynclinal sediments are underlain by platform or miogeosynclinal deposits.

(2) A broad, miogeosynclinal belt extends between the central eugeosyncline and the European foreland to the north and west. This belt covers essentially the Helvetic realm of the Swiss Alps and the Subalpine chains of France. It was invaded by the Flysch facies, which possibly should be classed as eugeosynclinal, during Late Cretaceous and early Tertiary time.

(3) Inside the eugeosyncline and south of it are unstable platforms, some of which are miogeosynclinal: the Briançonnais facies belt (broad sense) of the western Alps and parts of the Austroalpine nappes and southern Alps.

In other words the eugeosynclinal stage characteristically follows an interval of normal shallow-water sedimentation, which may or may not show an abnormally high rate of subsidence. During evolution of the geosyncline, eugeosynclinal conditions tend to spread out from the central troughs, where they originated in the Early Jurassic, and to encroach upon the adjoining platforms as well as over parts of the bordering miogeosyncline. This shifting is particularly pronounced during the later (Flysch) phases. On the other hand, no instances of a restoration of miogeosynclinal or platform conditions are known (in the Alps) once a eugeosynclinal belt has been established.

Triassic of the southern Alps.—The famous Triassic section of the southern Alps may serve as an example for the type of sedimentation that prevailed during the neutral phase between the Hercynian and Alpine orogenic cycles. Comprehensive summaries have been published by Leonardi (1955) on the Dolomites, and by de Sitter and de Sitter (1949) on the Bergamask Alps. Among the monographic descriptions may be mentioned Ogilvie-Gordon (1927), Pia (1937), E. Trümpy (1930), and Wirz (1945).

The facies development of the Permian and lowest Triassic indicates that geosynclinal evolution was in advance in the eastern area, in the Dolomite Mountains. With the early Middle Triassic Anisian stage, carbonate sedimentation becomes the rule. The thickness of Anisian limestones, dolomites, and (rare) bituminous shales, however, is still moderate (50–250 m).

The upper Middle Triassic or Ladinian, on the other hand, may exceed 1000 m and exhibits some striking facies changes. Three facies types can be distinguished:

(1) The so-called "normal" facies, consisting of shales with thin limestone bands and chert, tuffaceous sandstones, and, in the Dolomite Mountains, great masses of basic to intermediate volcanic rocks. Most of these are submarine lava flows or tuffs, but some may be near-surface sills. Even in Lombardy and in the western part of the northern calcareous Alps, thin tuffite members mark distant manifestations of these eruptions. In the Dolomites, the sea may have been fairly deep during deposition of the lower Ladinian, which contains an almost exclusively pelagic fauna dominated by radiolarians and the floating lamellibranch *Daonella*. The frequency of land plants in the middle Ladinian indicates that some volcanic cones rose above sea level. (A particular formation, the siliceous "splinter shale" of the Val di Scalve, in the Bergamask Alps (de Sitter,

[4] The palinspastic cross section given by Cadisch (1946, p. 11) is very impressive in this regard, but the thickness of the eugeosynclinal Bündnerschiefer formations has certainly been underestimated.

1949, p. 146), is important as it provides a link with the Partnach shales of the northern calcareous Alps.) The term "normal facies" for the shales with volcanic rocks is, of course, quite misleading, for the Middle Triassic outside of South Tyrol is generally represented by dolomites and limestones.

(2) Poorly bedded dolomites and limestones of the "reef" facies, the most famous of which are the Schlern (Sciliar) dolomites and the Marmolata limestones of the Dolomite Mountains, and the Esino limestones on Lake Como. Calcareous algae play a large part in the construction of these carbonate masses. Why some of them were dolomitized and others not is an unsolved problem.

(3) A euxenic facies of black, bituminous limestones with dolomite bands occurs in the upper Anisian and Ladinian on Lake Como and Lake Lugano. It contains some tuffite members as well as the fossil beds of Monte San Giorgio, with a rich and very well preserved fauna of marine reptiles but almost no bottom-dwelling organisms.

The limestone and dolomite areas show very sharp borders against those of the coeval shales and volcanic rocks, although the contact has in many instances been disturbed by later tectonic movements along these zones of mechanical discontinuity. Stratification in the carbonate rocks tends to slope outward near the borders, and tongues as well as isolated blocks of limestone penetrate the surrounding shale areas. During Ladinian time, the "reef" facies migrated outward into the hitherto volcanic basins. The limestone and dolomite masses are arranged in strips with a west-northwest to east-southeast trend. The contrast between the stupendous dolomite cliffs and the lush pasture lands on the volcanic rocks—both cut out of rocks of the same age—gives its unique character to the landscape of the Dolomite Mountains. M. Ogilvie-Gordon (1927, p. 122) proposed a very plausible explanation for these abrupt facies changes. She believes that the volcanic rocks were erupted from fissures in deeper-lying compartments, while carbonate deposition went on unhampered on the intervening platforms. The lime-secreting algae and corals could not thrive in the volcanic troughs because of the greater depth and the scarcity of oxygen. (*See* also Cornelius, 1925, p. 375.) It is probably not quite correct, however, to call the dolomite and limestone masses of the southern Alps bioherms, in spite of their great thickness and limited lateral extent.

During the Late Triassic, conditions were again more nearly uniform, although the Carnian still shows a variety of facies types. Thick dolomites and limestones of shallow-water origin are of Norian age.

The overall thickness of the Triassic formations reaches 3500 m in some areas of the southern Alps, the Triassic of the northern calcareous Alps attains 2000 m, and the Lower and Middle Triassic of the Penninic Briançonnais belt reaches 1000 to 1500 m. These formations may well be regarded as miogeosynclinal. Whether the Ladinian of the Dolomites, with its volcanic rocks and locally greater depth of the sea, marks a first abortive attempt toward the realization of eugeosynclinal conditions remains conjectural. There is some evidence for very slight tectonic disturbance between the Ladinian and Carnian stages, and we shall see that this very moment marks a turning point in Alpine history.

Lower Cretaceous of the Helvetic nappes.—Miogeosynclinal subsidence started in the Lias of the more southerly part of the Helvetic facies belt in the Swiss Alps and shifted to the site of the autochthonous Aar massif in the Late Jurassic. Most of the mountains in the domain of the Helvetic nappes are carved out of the thick and well-stratified, mainly calcareous formations of the Lower Cretaceous. Still the most comprehensive study of the rocks is Arnold Heim's classical monograph (1916); good general accounts are given by Albert Heim (1921) and Cadisch (1953).

In the cover of the Aar massif in eastern Switzerland, the Lower Cretaceous is only about 100 m thick, except where pre-Eocene erosion has removed it. It thickens rapidly southward and attains 1500 m in the higher Helvetic nappes; one stage alone, the Hauterivian, thickens from 20 to 600 m over a distance of 40 km across the strike. The predominant rocks are shallow-water limestones, such as rudist, crinoidal, bryozoan, oölitic, calcarenitic, and pellet limestones,[5] siliceous limestones, fine-grained limestones, calcareous shales, and glauconite beds with phosphatic fossil casts. Toward the south, most of these shallow-water limestones pass laterally into shales or into fine-grained pelagic limestones. (*See* palinspastic cross-section in Arnold Heim, 1916, p. 480.) This denotes a slope of the sea floor from the foreland margin toward the geosyncline. Still farther inward, in the Ultrahelvetic

[5] Most "onkoids" (Arnold Heim, 1916, p. 566) or "false oöliths" are faecal pellets (Brückner, 1951, p. 198).

realm, the Lower Cretaceous thins again and is almost entirely of pelagic facies ("Néocomien à Céphalopodes"). This announces the passage to conditions where subsidence exceeds sedimentation.

Rhythmic sedimentation is characteristic of the Helvetic Cretaceous (Arbenz, 1919; Fichter, 1934; Brückner, 1937). A typical cyclothem starts with a thin, condensed bed, generally rich in glauconite and pyrite, followed by shales, which grade upward into limestones. The top of the limestone members is sharp and represents a break in sedimentation. Diameter of detrital quartz grains is smallest n the shales, increases in the limestones, and reaches its maximum in the thin glauconite beds, whereafter it drops sharply. This type of sedimentary rhythm starts with the upper Sinemurian and has its most characteristic development in the Lower Cretaceous. The middle Cretaceous cyclothems are "degenerated;" several of them are crowded into a few meters. They disappear in the Upper Cretaceous, which is of pelagic facies. (*See* Bolli, 1944.) Such cyclothems are common in the shelf formation of the foreland, but they disappear toward the axis of the geosyncline, where sedimentation was more continuous.

A convincing interpretation of these rhythms yet remains to be found. Most authors agree that the gradation from shales to limestones indicates a gradual shallowing of the sea, be it by filling with sediment or by depression of the sea level. The glauconite beds were laid down under persistent conditions of inhibited sedimentation; in some instances ammonites of two or more zones are found in the same bed. Erosion is certainly submarine in most cases (Arnold Heim, 1925). It is hard to understand why sedimentation was slowed down during this phase, which should coincide with subsidence of the sea bottom. The role of compaction is certainly not to be neglected. Brückner (1951; 1953) has advanced an alternative hypothesis, relating the cyclothems to hydroclimatic fluctuations. This gives a plausible answer to some questions and raises others. Carozzi has published a great number of notes (summarized in his paper of 1951) showing that the rhythms were not so regular as often supposed; he has done very valuable work measuring the size and frequency of all the components in the sediments, and has shown enviable optimism in plotting these measurements in "curves," finally resulting in a "bathymetrical curve." For Carozzi, the rhythms of subsidence are related to tectonic movements. This seems rather improbable, as the regular cyclothems cannot be detected in the structurally more active belts of the geosyncline, whereas they are common far outside the Alps. Quite clearly, the cyclothems register some widely effective (world-wide?) pulsating mechanism, not directly related to the preparatory disturbances in the Alpine geosyncline.[6]

Jurassic of the Median Prealps.—Yet another tectonic environment of more or less miogeosynclinal characteristics is that of unstable platforms inside the eugeosynclinal belt, or between the central eugeosyncline and adjacent ordinary miogeosynclines. Sedimentation in this type of environment is exemplified by the Jurassic and Cretaceous series of the Median Prealps nappe (Fig. 2; Pl. 2). A summary of the stratigraphy of the Median Prealps has been given by Tercier (1952), and a very thorough review of older publications by Jeannet (in Albert Heim, 1922). The stratigraphy of the Subbriançonnais and Briançonnais belts of the French Alps, which continue the same facies zones southward, has been summarized by Gignoux (1950) and Moret (1954).

During the Early Jurassic, the belt of the Médianes plastiques (Peterhans, 1926) and of the analogous Subbriançonnais proper (Schneegans, 1938; Debelmas, 1955), which was only about 30 km broad, was divided into two or three minor troughs and submarine swells. Sedimentation in the former was fairly continuous and produced impure, cherty limestones, followed by silty, calcareous shales. Along the swells, crinoidal limestones predominate, and the ancient substrate was commonly indurated ("hardground") and pierced with annelid borings, where sedimentation was interrupted for the duration of a stage or longer. The southeasterly part of the Préalpes Médianes (Médianes rigides) and the Briançonnais

[6] A possible mechanism was suggested to the author by Preston Cloud while this paper was in preparation. Cloud asks: "Is it possible that cyclothemic sedimentation on the shelves is related to episodic large-scale dumping by gravity mass movement of sediments from the trough-margins (or heads) onto trough-bottoms? Isostatic adjustment to such abrupt transfer of load would logically result in local upwarping of the shelves, concurrently with local acceleration of trough subsidence. Repeated small upward movements in a shelf area undergoing regional subsidence could give rise to a succession of cyclothemic deposits; and the mechanism is consistent with, if not a necessary corollary of, the observed redeposited coarse clastics that are interbedded with the predominating pelagic shales of adjacent flysch basins."

proper lack Liassic rocks, although some Rhaetic and lower Lias was laid down and later eroded (before Bathonian time).

The facies distribution of the Middle Jurassic is often mentioned. To the northwest, we find the "*Zoophycos* Dogger," several hundred meters of calcareous shales alternating with argillaceous and silty limestones. The fauna consists mainly of cephalopods and of the brushlike dwellings called *Zoophycos* or *Cancellophycus*. As one approaches the southeastern border of this trough, oölitic and sandy limestones with drifted plant remains make their appearance; they lead to the much thinner "Intermediate Dogger," made up of oölitic and calcarenitic limestones. Still farther southeast, the Middle Jurassic is lacking, and finally one enters the domain of the "*Mytilus* beds," with brown, bituminous limestones and coal beds, containing a bottom-dwelling fauna with many endemic species but no cephalopods. At the base of the *Mytilus* beds, a laterite band has been found (Ellenberger, 1958, p. 293; Badoux and de Weisse, 1959), proving that the Briançonnais proper was emergent before the Bathonian.

In the Upper Jurassic, limestones are formed all over the Médianes area (for detailed analysis, *see* Weiss, 1949). The Lower Cretaceous pelagic, thin-bedded limestones with chert are again found only in the northwestern part, whereas the pelagic *Globotruncana* and *Globorotalia* limestones of the Couches Rouges (Cenomanian to Paleocene) spread out uniformly over the whole hitherto differentiated facies belt and neighboring realms.

Characteristic of this kind of facies belt is the abrupt vertical and horizontal change of environment. The thickness of the Middle Jurassic, for instance, jumps from nothing to 700 m within 4 km across the strike. Sediments of the same age, laid down a few kilometers apart, may show coal beds in one place, deep-water mud facies in another. Bathymetric interpretation is of course commonly hazardous, but it seems highly probable that the monotonous sequence of the *Zoophycos* beds was formed under several hundred meters of water. In spite of this, the facies belt as a whole generally rose above the adjoining eugeosynclinal realms and was not invaded by volcanic material. (This statement is made with some reservation

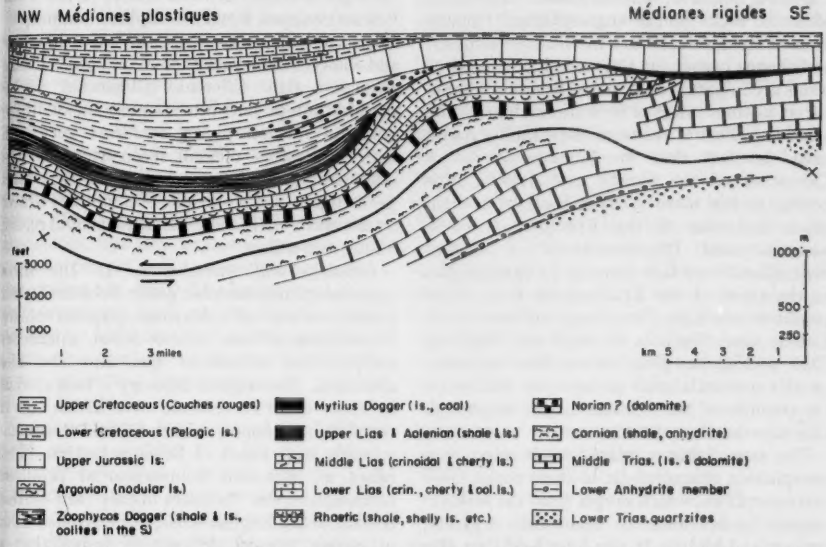


FIGURE 2.—PALINSPASTIC CROSS SECTION THROUGH THE MESOZOIC FORMATIONS OF THE MÉDIANES NAPPE IN WESTERN SWITZERLAND

Vertical scale exaggerated × 4. After Gagnebin, Jeannet, Lugeon, etc. (references cited)

The Triassic formations shown below the basal thrust plane of the nappe have not been stripped from their Paleozoic basement and are now found on the front of the Grand St. Bernard nappe, in Valais.

as far as the relation between the Subbriançonnais and Valais belts during the early and middle Jurassic is concerned.)

Bündnerschiefer, Schistes Lustrés

General statement; metamorphism.—The usual Mesozoic rocks of the eugeosynclinal troughs of the Penninic realm are the Bündnerschiefer, or Schistes lustrés, or Calcescisti. These three terms are in part synonymous; Bündnerschiefer covers a wider range of formations than do the other two. Part of the misunderstanding comes from the fact that the Schistes lustrés or Calcescisti of the French and Italian Alps belong to the more internal Piemont trough, south and east of the Briançonnais swell, whereas most Bündnerschiefer of eastern Switzerland were laid down in a eugeosyncline to the north of the Briançonnais belt, which persisted into Late Cretaceous and even early Eocene time. Nabholz (*in* Bolli and Nabholz, 1959, p. 238) has given an excellent summary on this nomenclatorial question, as well as on the problem of the Schistes lustrés in general. (*See* also dal Piaz, 1943.)

The term Schistes lustrés[7] itself refers to a conspicuous characteristic of these rocks, their metamorphism, which ranges from the sericite-chlorite zone down to staurotide, kyanite, garnet, and biotite. It was long held that this metamorphism was of primary, geosynclinal nature and inherent to the Schistes lustrés facies. Recent research (*e.g.*, Ellenberger, 1958; E. Niggli, *in* Cadisch 1953; Wenk, 1956) has discredited this view, which was already opposed by Cornelius (1930). Alpine metamorphism is now proved to have occurred late in the history of the mountain chain, not before the end of the Eocene (Ellenberger, 1952a). This leaves only the Oligocene and part of the Miocene for the development of metamorphism. In a given area, it has affected the (essentially noncalcareous) pre-Triassic basement in the same degree as the Mesozoic rocks. In the Vanoise Mountains of Savoy, the platform sediments of the Briançonnais belt are just as metamorphic as the Schistes lustrés, and in the northern Ticino mesozonal metamorphism has transformed the miogeosynclinal Liassic formations in the cover of the Gotthard massif (± Ultrahelvetic) to about the same extent as the eugeosynclinal Schistes lustrés of the lower Penninic nappes. Breaks in Alpine metamorphic zones are found only where the latest Miocene and Pliocene movements have brought tectonic elements with different (Oligocene) metamorphic antecedents into contact. Incidentally, grade of metamorphism conforms very well with the classical tectonic reconstructions, and isometamorphic surfaces are very nearly parallel to the top of the Alpine pile of nappes as drawn on the much-criticized cross sections by Argand, Staub, and others.

Lithology and environment.—In the most common epimetamorphic grade, Schistes lustrés consist mainly of calcareous sericite schists. Noncalcareous slate, sericite schist, micaceous marble, and micaceous quartzite are also abundant. The original lithology is best studied in areas of little tectonic overburden, or in décollement nappes, which contain nonmetamorphic equivalents of Schistes lustrés. (This raises an awkward nomenclatorial problem: nonmetamorphic "Schistes lustrés" are neither schists, in the English sense, nor lustrous.) It is, of course, always difficult to prove that a formation really can be regarded as nonmetamorphic Schistes lustrés as long as the relationship of the cover nappe with the central part of the Alps is controversial. It seems likely, however, that the Liassic "Schistes inférieurs" of the Breccia nappe, the Jurassic of the Grande Moënda digitation in the Pas du Roc nappe (Barbier, 1948, p. 49), or the Jurassic and Cretaceous shales of the Arosa zone may give a fair idea of what Schistes lustrés looked like before sedimentary structure was obscured by metamorphism.

Thin-bedded, argillaceous, silty or more rarely sandy limestones, alternating with calcareous shales, presumably were the predominant rocks of the original Schistes lustrés. Marly, argillaceous or silty shales form the major part of some formations. Sandstones are fine-grained (diameter of most quartz grains 0.02 to 0.2 mm), commonly argillaceous or calcareous, and do not contain much feldspar. Many of the limestones are calcarenitic. Microbreccias, mainly with angular fragments of Triassic limestones and dolomites, occur in several areas; they generally show graded bedding, whereas most of the sandstones do

[7] Stratigraphical analysis of the Schistes lustrés in the broad sense is still in its beginning. Working in these areas is like trying to solve one equation with two variables, stratigraphy and structure (W. Leupold, orally); a solution is possible only by making a reasonable assumption for one of them, say that a certain sequence of formations represents a normal stratigraphic column. The author recognizes from bitter experience that a new fact, such as an "unlucky" fossil or some graded beds showing inversion, can bring the most beautiful working hypothesis tumbling down.

not. Carbonate, quartz sand, and argillaceous matter are intimately mixed. Pyrite is abundant, but euxenic bottom conditions do not appear to have been the rule.

The individual formations of the Schistes lustrés group are ill-defined and grade into one another. Nothing is left of the regular cyclothems, which render the lithostratigraphic subdivision of the miogeosynclinal sediments in the Helvetic belt so clear cut. Lateral facies changes are common, but research has not progressed far enough to allow detailed analysis.

The fauna (Bolli and Nabholz, 1959) is extremely poor. This sterility seems to be an original feature, although metamorphism may have destroyed many fossils. (*See* discussion in Bucher, 1953.) Radiolarians are most frequently encountered, but tiny mollusca, bryozoan and crinoid fragments, and rare, mainly arenaceous Foraminifera also occur. The few larger shallow-water organisms, such as the Liassic oysters described by Nabholz (1945, p. 43), were found in beds near the base of the group, which do not yet show the characteristics of the real Schistes lustrés.

We may assume that most of the Schistes lustrés were laid down in fairly deep water, because of both lack of typical shallow-water sediments and fauna. The environment suggests only moderately steep submarine slopes, save for the belts in which breccias were formed. Turbidity currents still are rather exceptional. The supply of detrital matter, derived chiefly from Triassic and Liassic rocks, was sufficient to allow the deposition of the thick Schistes lustrés group, during a time span which seems to have covered a few tens of millions of years in each eugeosynclinal belt. The Schistes lustrés of the Piemont belt reach about 1500 or 2000 m, but the later "pre-Flysch" formations of the Valais belt may be much thicker. Evaluation of original thickness is, of course, very problematic in the central zone of the Alps.

Many geologists believe graywackes are *the* typical rock of eugeosynclinal facies assemblages. The author has never seen a bed of graywacke in the Alpine Schistes lustrés. It is rather an essentially calcareous (but nowhere dolomitic) rock group. This is a major difference from many older, Paleozoic and Precambrian eugeosynclines, in which graywackes and non-calcareous shales are predominant. This may be due to the generally warm hydroclimate of the Mesozoic Tethys seas, or it may simply indicate the difference in geological age, as there was no calcareous microplankton in the Paleozoic.

Ophiolites (Roches vertes, Pietre verdi) are commonly associated with Schistes lustrés, although there are otherwise quite normal Schistes lustrés formations without volcanic rocks. The term has an extremely broad meaning, covering ultrabasic and basic, extrusive and intrusive rocks: peridotites, gabbros, porphyrites, spilites, tuffs, and their various metamorphic derivates such as serpentines, amphibolites, prasinites, ovardites, and other "greenstones" (Burri and Niggli, 1945). The extrusive nature of these magmatic rocks is in many cases attested by pillow lavas and tuffitic beds. The lava outflows seem to have been entirely submarine; there is nothing to prove the existence of a belt of volcanic islands in the Mesozoic history of the Alpine geosyncline. It is possible, but not certain, that andesitic volcanoes existed at a much later time, at the end of the Eocene. Other sheets are probably sills and laccolites. All authors seem to agree that most ophiolites are of Cretaceous, especially Early and middle Cretaceous, age, but some peridotites may be older. The hypothesis of synorogenic intrusion along early thrust planes seems to be constantly losing ground in favor of pre-orogenic submarine flows and intrusions into unconsolidated, horizontal sediments.

Schistes lustrés of the Cottic Alps.—The Cottic Alps, in southwestern Piemont and adjoining areas of the French Durance basin, are the region where stratigraphic analysis of the Schistes lustrés in the internal, Piemont eugeosyncline has the greatest chance of success. Metamorphism is much less pronounced than farther north, probably because the great overburden of the Dentblanche nappe was no longer present, and the structure is easier to work out. Even here, however, many questions are still moot. The most important older works are numerous papers by Franchi, between 1898 and 1929. Recent work by Conti (1953; 1955) and Lemoine (1951; 1954; 1955; 1957; 1959) has greatly advanced our knowledge.

The most significant section is the one studied by Lemoine (1951) on Mount Gondran, east of Briançon (Fig. 3). Subsidence in this part of the Piemont trough started in the Carnian, but normal shallow-water deposits with benthonic fossils were laid down well up into the Hettangian. Dark, cherty limestones form a transition from these neritic sediments to typical, although only slightly metamorphic, Schistes lustrés, with thin, graded beds of microbreccia. The upper part of this formation, the Gondran "Flysch," contains some sandstones. It is overlain by thin ("leptogeosynclinal") beds of

radiolarian chert and fine-grained, maiolica-type limestones which elsewhere in the Alps and in the Apennines represent the upper part of of the Middle Jurassic, the Upper Jurassic, and the base of the Cretaceous. The ophiolites,

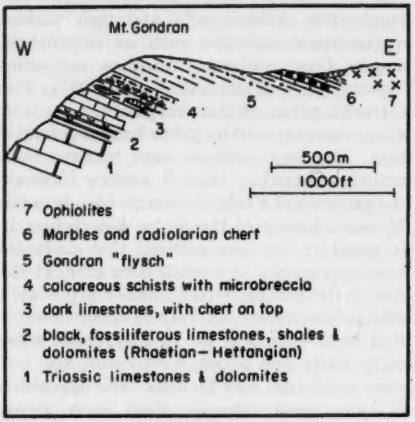


FIGURE 3.—SECTION AT MOUNT GONDRAN, EAST OF BRIANÇON
After Lemoine (1951; 1955)

which appear on top, are submarine lava flows, almost certainly of early Cretaceous age.

Mount Gondran belongs to an external digitation of the Schistes lustrés nappe, which overrides the Briançonnais zone; but quite similar sections are reported by Lemoine himself and by Conti from farther east. There, the Schistes lustrés proper are much thicker than the Gondran "Flysch" (about 2000 m?) and contain less microbreccia. Below them, Norian to Sinemurian formations are dated by fossils. A part of the ophiolites, commonly associated with cherts and maiolica limestones, seem to form synclines on the mass of Schistes lustrés, so that it is probable, if not certain, that the Schistes lustrés of the Cottic Alps are of pre-radiolaritic age, and that they were formed during the middle and late Lias and the earlier Middle Jurassic.

Recently, however, and in the light of new field observations, Lemoine (1959, and personal communication) has come to doubt his original interpretation, as here reported. He now suspects that the Gondran "Flysch" does not at all belong to the same unit as the overlying cherts, limestones, and ophiolites, and that it may be much younger than the undoubtedly Liassic strata on which it rests. The Schistes lustrés of the Val Grana, on the Italian side, are also separated from the fossiliferous Triassic and lower Lias in their substratum, by a thrust plane marked by a thin seam of Triassic cornieule. Lemoine thinks it possible that the ophiolites in "synclinal" position on the Schistes lustrés, together with the accompanying cherts and limestones, might be relics of a higher "Ophiolitic nappe," unrelated to the Schistes lustrés proper, like the upper part of the Gondran section (Nos. 6 and 7 of Fig. 3). In this hypothesis, there is no more direct proof that the Schistes lustrés are really pre-radiolaritic, and they might even be of Cretaceous age. This would make them coeval with the mainly Cretaceous Bündnerschiefer of the external Valais eugeosyncline; Lemoine rightly holds that a comparison between the stratigraphic evolution of these two basins is premature (personal communication).

Publication of Lemoine's new results is awaited eagerly, but the author admits that he found the earlier hypothesis rather satisfying.

According to Franchi and Conti, Schistes lustrés sedimentation started as early as the Late Triassic in the central part of the southern Piemont geosyncline. Below limestones, which are thought to be of Rhaetian and Liassic age, there is a thick formation of phyllitic schists with beds of calcite and dolomite marbles and sodic volcanic rocks which Conti regards as a time equivalent of the Norian Hauptdolomite formation farther east. But the tectonic and stratigraphic relationship of this "Montenotte formation" to the Schistes lustrés is rather obscure. At any rate, no Triassic Schistes lustrés are known from other regions of the Alps.

Bündnerschiefer of Graubünden.—Large tracts in western, central, and northern Graubünden consist of the monotonous Bündnerschiefer group. These sedimentary rocks belong to the sheared-off Mesozoic cover of the lower Penninic nappes, especially the Adula gneiss nappe. An excellent description of them was given long ago by Albert Heim (1891). Stratigraphic subdivision was attempted by Staub and especially by his pupils H. Jäckli (1941) and W. Nabholz (1945; 1951).

The most detailed section is that of the highest (Tomül) strip sheet, roughly summarized in Figure 4. One observes at first glance the difference between these Bündnerschiefer of the external Valais eugeosyncline and the Piemont Schistes lustrés. The only significant Bündnerschiefer fossils so far found are lower Liassic oysters (*Gryphaea*). They

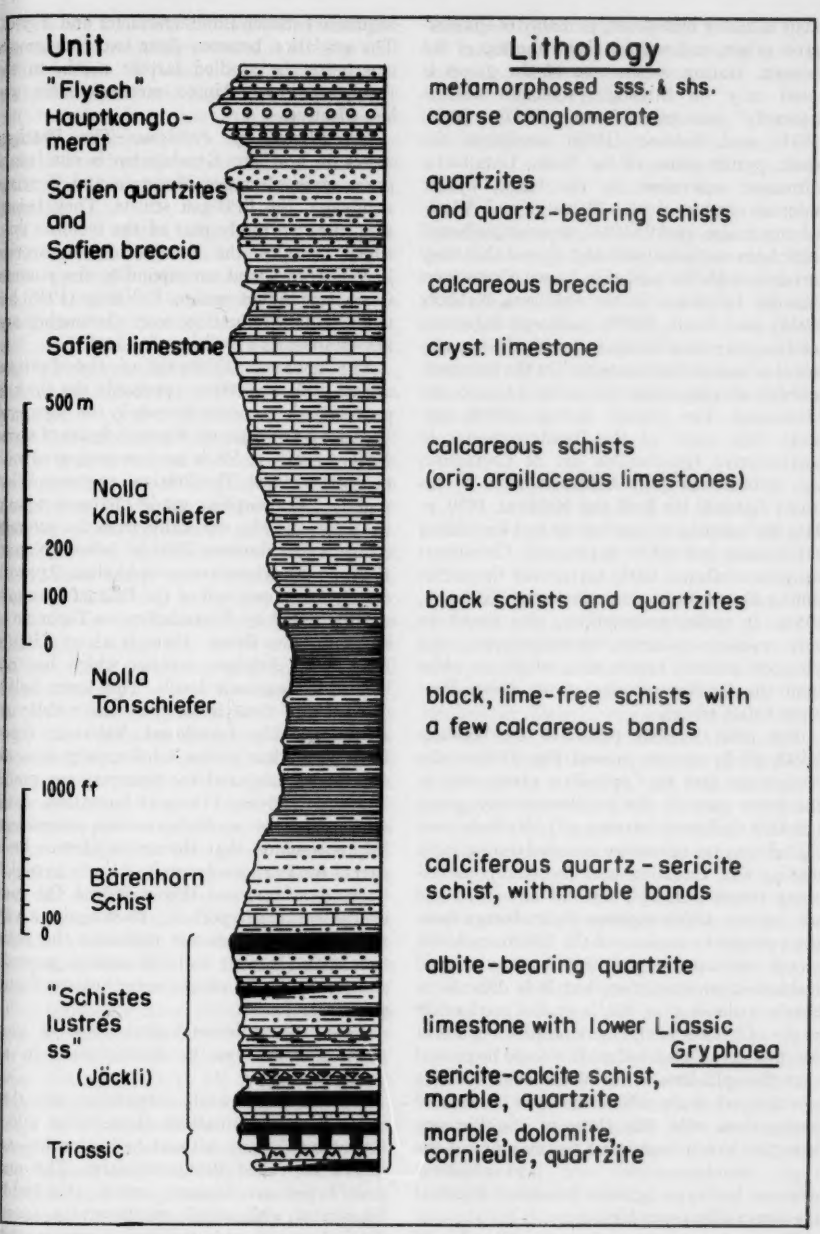


FIGURE 4.—SECTION OF PART OF THE BÜNDNERSCHIEFER IN THE TOMÜL SHEET OF WESTERN GRAUBÜNDEN
After Jäckli (1941) and Nabholz (1945). Ophiolites (mainly sills) are shown in black

occur in sandy limestones, probably of shallow-water origin, and not far above the top of the Triassic. Dating of the rest of the group is based only on lithologic analogies and is extremely uncertain. Staub (1937), Jäckli (1941), and Nabholz (1945) considered the black, pyritic slates of the Nolla Tonschiefer formation equivalent to the black, pyritic Aalenian shales of the Helvetic and Ultrahelvetic realm. Jäckli (1944), however, followed these beds northeastward and proved that they correlate with the probably Lower Cretaceous Valzeina formation in the Prätigau. Nabholz (1951) and Staub (1958) proposed ingenious solutions to reconcile this fact with the hypothesis of a "comprehensive series" in the Bündnerschiefer, encompassing the entire Jurassic and Cretaceous. The present author (1958) suspects that most of the Bundnerschiefer in northwestern Graubünden are of Cretaceous age, but he was quite rightly rebuked by his friend Nabholz (*in* Bolli and Nabholz, 1959, p. 241) for jumping to conclusions and for adding yet another unverified hypothesis. Cretaceous age is nevertheless fairly certain for the rather similar Ferret schist group in Valais (Trümpy, 1954). If earlier assumptions, also based on very meager evidence, prove correct, the Piemont Schistes lustrés as a whole are older than the Bündnerschiefer of the North-Penninic Valais trough.

The cross sections published by Nabholz (1945, Pl. 5; see also present Fig. 4) show the remarkable fact that ophiolites occur only in the lower part of the Bündnerschiefer group (Jäckli's "Schistes lustrés s.s."). Nabholz concluded that the ophiolites intruded during early folding and thrusting and commonly moved along thrust planes. They did not reach the sea bottom, which explains their absence from the younger formations of the Bündnerschiefer group. Schoenenberg (1956) has questioned Nabholz's interpretation, but it is difficult to obtain a decision on the intrusion mechanism of the ophiolites in the metamorphic regions of southwestern Graubünden. If it could be proved that the ophiolites in the Valais eugeosyncline are derived from submarine lava flows and near-surface sills, like those of the Piemont belt, this would imply that the majority of the type Bündnerschiefer are post-ophiolitic, whereas the type Schistes lustrés of Piemont are essentially pre-ophiolitic.

The youngest formations in the Tomül sheet, from the Safien limestone upward, to which all authors now assign a Cretaceous age, show the characteristics of an intermediate sequence between Bündnerschiefer and Flysch. The gradation between these two rock groups can better be studied farther north, in the Prätigau, where Alpine metamorphism was less effective.

"Pre-Flysch" of Prätigau.—The Prätigau valley in northern Graubünden is cut into a great mass of upper Mesozoic and Tertiary sediments, the Prätigau schists. They belong to a more northerly part of the tectonic unit, which contains the Bündnerschiefer section just described, and correspond to the younger formations of that section. P. Nänny (1948) has studied these formations very thoroughly and has established a well-founded scale.

The apparent thickness of the Prätigau schists is about 6000 m—probably the thickest post-Triassic sequence known in the Alps, and the base is not exposed. Nänny's figure of about 4000 m (1948, p. 59) is too low compared with his cross sections. The Prätigau sequence is but slightly metamorphic; grade increases toward the south and also vertically, from the summits to the valley bottom 2000 m below (Nänny, 1948, p. 46). There are no ophiolites. Approximately the upper half of the Prätigau group is reliably dated by Foraminifera as Turonian to lower Eocene. Below, there is about 2500 or 3000 m of Prätigau schists, which has not furnished diagnostic fossils. This lower half is divided into three formations; the middle one is the already mentioned Valzeina slates. Each formation grades lithologically into the next higher one, and the youngest one grades into the overlying Pfävigrat formation, dated as Turonian by its *Globotruncana* assemblage. Nänny assumes that the unfossiliferous lower part of the Prätigau group is of Early to middle Cretaceous age, and this is indeed the most logical working hypothesis. Here again, a very thick sequence does not represent the entire time span (Jurassic and Cretaceous) generally allotted to the Bündnerschiefer-Schistes lustrés group.

From the sedimentological point of view, three subunits can be distinguished in the Prätigau schists:

(1) The lower unit, comprising the three unfossiliferous formations (Lower and middle Cretaceous?), may without hesitation be—and indeed is—called Bündnerschiefer. The main rock types are impure, rather thin-bedded limestones with small quartz grains (partly detrital, partly recrystallized cherty matter), calcareous shales, and argillaceous shales (or rather slates) with some fine-grained calcareous sandstones. Thin layers of micro-

breccia are common but not prominent. The sequence is very monotonous, and rhythmic sedimentation, if present, is of the symmetric shale–limestone–shale type. Graded bedding is not recorded from the Prätigau, but some graded sandstones and microbreccias have been observed at equivalent levels of the Ferret schists, in Valais (P. Fricker, doctorate thesis, in press).

(2) In the second subgroup, of Late Cretaceous age, sedimentation is much more varied. It is essentially calcareous; sandy, more or less calcarenitic, limestones pass into calcareous sandstones, argillaceous and silty limestones, and marl slates. Breccias and conglomerates with poorly rounded pebbles are conspicuous, and the breccias in the older formations contain a greater proportion of elements from the basement than do the younger ones. The detrital matter is derived from the north. The breccias and microbreccias show graded bedding, but the grading of the finer-grained sandstones is generally poor. Cyclothems a few meters thick, comparable to those described by Lombard (1949) from the Niesen Flysch, are common. The various bedding phenomena characteristic of Flysch are incompletely developed. On the whole the sea appears to have been rather shallower than during deposition of the typical Bündnerschiefer, submarine slopes steeper, and tectonic activity stronger.

(3) The Paleocene and lower Eocene formations constitute the third unit. It is a typical Flysch, with argillaceous shales, quartzitic sandstones, and coarse, graded, feldspathic sandstones, poor in carbonate.

The Prätigau schist group is especially important, as it shows the vertical gradation from the Bündnerschiefer type of sedimentation to that of Flysch. This gradation is also lateral; proceeding southward the Upper Cretaceous "pre-Flysch" formations become finer in grain, more uniform in character, and indistinguishable from Bündnerschiefer. It is quite impossible to draw a sharp boundary line between the two groups of sedimentary rocks. Nänny (1948, p. 64) proposes to lump everything together under the common denominator of Bündnerschiefer. This is certainly going too far. In other mountain chains (*e.g.*, the Carpathians) the opposite solution is adopted, and everything is called Flysch. Bertrand (1897) distinguished the "faciès Flysch shisteux" (Schistes lustrés etc.) and the "faciès Flysch grossier" (Flysch in the present Alpine sense). The "Flysch with ophiolites", observed in many fold belts, would generally correspond more nearly to the Schistes lustrés of the Alps, there being only very rare and generally dubious occurrences of volcanic rocks in typical Flysch.

In their typical development, Bündnerschiefer (including Schistes lustrés) and Flysch are quite distinct rock groups, but there are many transitions between the two. The Bündnerschiefer are restricted to the eugeosynclinal troughs, where they are commonly associated with ophiolitic volcanic rocks. Their sedimentation precedes the folding movements. The Flysch facies, which will be discussed more fully, marks the "filling-up" of the geosyncline and its encroachment onto areas that had hitherto shown platform or miogeosynclinal characteristics, concurrently with folding and thrust faulting.

"*Leptogeosynclinal*" *Formations*

Meaning of the term.—The term "leptogeosynclinal" (from *leptos*, thin) was proposed by Trümpy (1955, p. 220) for thin deep-sea deposits representing a long time span. This distinction was believed necessary as long as many geologists held that geosynclines were always characterized by abnormally great thickness of their sedimentary filling, as indeed implied by Dana's original definition. Alpine geologists, in the wake of Haug (1900), believe that sedimentation in true eugeosynclines could also be "starved," as long as, for one reason or another, little detrital matter is furnished and the growth of shell- or skeleton-bearing bottom organisms is hampered. The deep sea is the most common, but by no means the only, environment which can cause extreme slowing down of sedimentation. (*See* also Sonder, 1939.) Today, many workers agree that eugeosynclines are not necessarily the sites of abnormally rapid sedimentation. (*Compare* Kündig, 1959.) If this agreement were to become general, the term "leptogeosyncline" could easily be spared; it was half seriously introduced to stress the difference between Dana's and Haug's conceptions of geosynclines, which might have been called, 10 years ago, the "American" and "European" concepts.

Recognition of the leptogeosynclinal nature of a formation depends, of course, on the evidence that permits the inference that a certain kind of sedimentary rock has been formed at bathyal or greater depths. Much of the evidence relating to this question is negative, and none of it,

taken alone, suffices to prove the deep-sea origin of a formation. But the concurrence of several of the proposed criteria lends a reasonable probability to the idea that many Alpine sediments have indeed been laid down several kilometers below sea level, and under conditions approaching those in the present oceans.

Radiolarian cherts.—Red, or more rarely green or black, thin-bedded cherts with Radiolaria are found in the upper Middle Jurassic and lower Upper Jurassic of the Penninic Piemont eugeosyncline and in all the units of equivalent age south and east of it, down to the southern Alps. They occur all over the Mediterranean, in Greece, in Italy, in southern Spain, and in northwestern Africa. Manganese compounds, as coatings or more rarely as nodules, are common. Grunau (1957) found minute particles of ultrabasic rock in certain green radiolarites. The cherts themselves, which in some instances contain but few radiolarians, are generally free of carbonate, but they may be interbedded with very fine-grained limestones. Grunau (1959) has just published a thorough review on the problems of the radiolarian cherts of the Alps and associated rock types, and Cornelius (1951) has also discussed the question in a very clear and unbiased manner.

The bathymetric significance of the radiolarian cherts has been the object of many discussions since Steinmann (1905; 1925, foreshadowed by Wähner, 1903) in his brilliant papers compared them to recent radiolarian oozes and interpreted them as abyssal sediments. In the 1930's and 1940's, European geologists tended to consider practically all visible sedimentary rocks as having been formed in shallow water, and the radiolarites suffered the same fate (*see* especially Grunau, 1947, p. 37–53) even though Tercier (1939, p. 77) opposed these too schematic views. Criticism of Steinmann's concept was based mainly on the association of radiolarian cherts with supposed shallow-water formations, such as breccias and sandstones. The origin of the marine breccias in the Alps will be examined later. The sandstones apparently all are turbidites, characteristic of deep rather than shallow water. Other arguments concern the chemical composition of the cherts, much richer in silica than are the present-day radiolarian oozes. But there has certainly been much diagenetic transfer of silica, explaining perhaps also the segregation into pure cherts on one hand and limestones, with calcified radiolarian skeletons, on the other. The fact that some radiolarites conformably overlie shallow-water formations is not surprising, if one considers the enormous vertical movements during short time spans brought to light by modern oceanography. Silicified driftwood, found in some places, has no bathymetric significance.

The frequent association of radiolarian cherts with ophiolites has been noted by all observers since Steinmann. For Wenk (1949) and others, this association is significant, and the proliferation of Radiolaria is supposed to be due to the freeing of silica on the sea floor during the spilite reaction. But there are many occurrences of radiolarian cherts far from any manifestations of submarine volcanism. In other cases, the ophiolites can be proved to be younger than the radiolarites, and hence fail as a source of silica. At any rate, it will not do to assume a volcanic supply of silica for the cherts in the Arosa zone, associated with ophiolites, and another (unknown) source of silica for those in the southern Alps (Grunau, 1947; 1959), as the same radiolarite formation extends, practically unbroken, over the whole intervening Austroalpine realm and far beyond.

Taking into account the absence of benthonic fauna from radiolarian cherts, the presence of manganese nodules, the very slow rate of sedimentation–10 to 40 m of radiolarite can represent the time span from Bajocian to early Tithonian, without any evidence of gaps—, and the vast horizontal extent of the radiolarites of equivalent age, there seems little reason to reject Steinmann's views on the deep-sea origin of radiolarian cherts in the Alps.

Pelagic limestones and azoic shales.—The limestones, which accompany the radiolarian cherts and overlie them in most sections, are extremely fine-grained, commonly cherty, and show stylolitic partings, which can crowd together to give a nodular structure to the rock. They contain pelagic microfossils only, such as tintinnids, small globigerinids, and the problematical *Nannoconus*. Ammonites are found locally, but in most places they are represented only by their calcitic aptychi; their aragonitic shells have been dissolved. These limestones (Maiolica, Biancone, Aptychenkalk, Hyänenmarmor, etc.) have apparently been laid down under conditions where calcite was stable, but no aragonite. Actual precipitation of the calcite, inasmuch as it does not belong to pelagic microfossils, may have taken place near the water surface. The thickness of these limestones is generally modest, although rather higher than that of the cherts. It appears that deposition was frequently interrupted.

The argillaceous or siliceous shales which

occur with the radiolarites and the maiolica limestones are always devoid of benthonic fossils, except for fucoids. This absence of remains of shell-bearing, bottom-dwelling animals not only in the cherts but also in the other rock types of the radiolarite association is certainly significant.

G. Colom (1957), who probably knows the pelagic formations of the Mesozoic and their microfauna better than anybody else, comes to the startling but well-founded conclusion that almost oceanic conditions prevailed over a large part of the country surrounding the western Mediterranean, during the Middle and Late Jurassic and the Early Cretaceous. This concept will certainly call for a thorough revision of current paleogeographical views. The Tethys Sea of middle Mesozoic time, often considered a narrow and comparatively shallow stretch of sea between two continents, probably is to be likened to a broad expanse of deep sea, with localized abyssal troughs—something, perhaps, like the existing Carribbean.

As far as the Alps are concerned the evidence may be interpreted as suggesting the existence of a broad expanse of deep sea, covering the Piemont, Austroalpine, and south Alpine realms during Middle Jurassic to Early Cretaceous time. A few islands or swells bearing shallow-water sediments (such as the Upper Jurassic Plassen limestone of the Hallstatt nappe) rose above the surrounding, paraoceanic bottoms. If the Préalpes Médianes-Sulzfluh nappe is of Penninic, Briançonnais origin, as the present author believes, the Upper Jurassic shallow-water limestones of this unit were laid down outside (north and west of) this deep-water realm. If, on the contrary, this nappe belonged to the lower Austroalpine units, the limestones necessarily formed on a broad rise surrounded by deep sea. The Upper Jurassic formations unequivocally assigned to the lower Austroalpine nappes of the Upper Engadine are entirely in a leptogeosynclinal facies of radiolarian cherts and maiolica limestones ("hyena marbles").

Jurassic and Cretaceous of Arosa zone.—The Arosa Schuppenzone of the Rhaetic Alps, one of the most interesting although very much sliced-up cover nappes of the Alps, lies immediately below the great Austroalpine thrust mass and above the Sulzfluh (Médianes) nappe. Several detailed studies have been devoted to it (Cadisch, 1921; Grunau, 1947; 1959; Gees, 1955; D. Richter, 1957). In this paper it is regarded as the highest Penninic unit; it can probably be correlated with the Platta nappe of southern Graubünden, with the wrongly so-called "Simme nappe" of the Iberg klippen on Lake Lucerne (Jeannet, 1941), and with the ophiolite-bearing nappe in the Prealps (Jaffé, 1955). Many authors, however, still consider the Arosa zone as a lower Austroalpine unit, in spite of its close relationship with the undoubtedly Penninic Platta nappe. The correlation here proposed does away with the highly anomalous occurrence of ophiolites in an "Austroalpine" unit, and ophiolites become an "index" to the eugeosynclinal Penninic realm.

Up to the Rhaetic or to the early Lias, the sea remained shallow, as attested by dolomites and limestones with benthonic fossils. These formations are followed by dark azoic shales with thin bands and lenses of finer-grained, dull-gray limestone. They show a curious resemblance to the (presumably younger) Argille Scagliose shales and Palombino limestones of the Liguride slide sheets in the northern Apennines. Any geologist would call this shale sequence Schistes lustrés, if it were metamorphic. Sedimentary breccias occur especially in its upper part, which is also richer in limestones. Five to 20 m of thin-bedded, red and green radiolarian cherts comes next. (For their detailed description, *see* Grunau, 1959, p. 79.) The overlying Aptychus limestone is dated as latest Jurassic and earliest Cretaceous by its pelagic microfauna. The Early Cretaceous is represented by shales with some graded beds of sandstone and breccia. The much-disputed Maran breccia, with angular fragments of dolomite, siliceous shale, and chert, is considered as syn-radiolaritic by Cadisch and Richter, as post-radiolaritic by Grunau and Gees. Locally, there are Barremian-Aptian limestones with dolomite fragments, overlain by Middle Cretaceous shales and greenish sandstones, which are quite similar to the Tristelkalk and "Gault" formations of the more external Falknis nappe.

The entire sequence between the lower Lias and the topmost Cretaceous contains essentially no benthonic fossils, although a few Middle Jurassic brachiopods have been found in one locality (D. Richter, 1957, p. 328) in a kind of rock which is highly unusual for the Arosa zone. The facies of the Jurassic and Cretaceous deposits can probably be attributed to deep-water origin of the sediments. The composite thickness rarely exceeds 150 m, most of which is taken up by the "nonmetamorphic Schistes lustrés" at the base.

In addition to the sediments there are large masses of ophiolites which predominate in the southern prolongation of the Arosa zone, the Platta nappe. Grunau (1957) advances

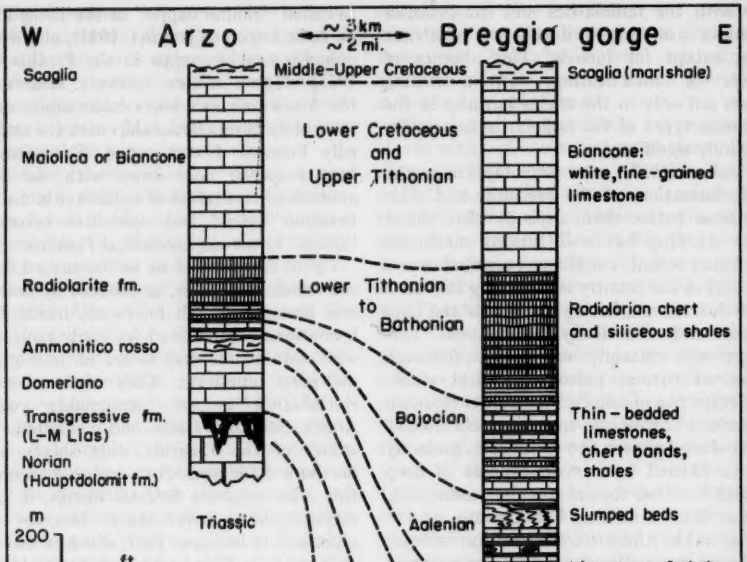


FIGURE 5.—SECTION OF PART OF THE JURASSIC AND CRETACEOUS IN SOUTHERN TICINO
After Frauenfelder (1916), Grunau (1959), Renz (1920), Senn (1924), and Vonderschmitt (1940). The two sections are only about 3 km apart; between them runs the Lugano fault.

good arguments for assigning these ophiolites to two phases: intrusive peridotites of Late Jurassic age and extrusive spilites, which flowed out in the middle and Late Cretaceous.

The Mesozoic series of the Arosa zone affords a fair example of a "lepto-eugeosynclinal" facies group. But the study of similar rock assemblages is much easier in other mountain chains, where the structural complications in the eugeosynclinal belt are not so enormous.

Jurassic and Cretaceous of the southern Alps.—The "leptogeosynclinal" deep-sea facies is not restricted to the eugeosynclinal belt but also occurs over a very wide area south and east of the Penninic realm. The well-known section in the Breggia gorge, in southern Ticino (Fig. 5, right-hand column) may serve as an example for the facies of the Jurassic in the southern Alps. From the middle Lias onward the facies is uniform all over the southern

Alps, except for the existence of another formation of red nodular limestones, the Ammonitico rosso superiore, of Kimmeridgian age, in the eastern part.

Above Norian dolomites and Rhaetic shales and limestones, the Jurassic starts with a great thickness (up to 1200 m) of regularly bedded limestones, with layers and nodules of chert, and some intercalations of crinoidal limestones. The fauna is poor, but sufficient to show the presence of all stages of the lower Lias and of the Pliensbachian (lower middle Lias). The Domerian, Toarcian, and Aalenian limestones, on the contrary, are very rich in ammonites, especially of the two pelagic groups of Phylloceratacea and Lytoceratacea, to the near exclusion of other fossils, except for planktonic or pseudoplanktonic lamellibranchs of the genus *Posidonia*. A slumped bed 12 m thick occurs in the upper Aalenian of the Breggia gorge, and several features in the Jurassic sediments of the surrounding area can probably be ascribed to syngenetic slumping. (*See* Grunau, 1959, Figs. 37, 39). From the Bajocian onward, the macrofauna almost disappears, apparently because of dissolution of aragonitic remains on the bottom; aptychi of ammonites are the only megafossils. The whole sequence clearly denotes a gradual deepening of the sea, which reached its maximum in the late Middle and early Late Jurassic when the radiolarian cherts were laid down. As long as the sea remained shallow, up to the middle Lias, sedimentation kept up with subsidence (miogeosynclinal conditions); after this, "starvation" set in, and the rate of sedimentation dropped sharply ("leptogeosynclinal" conditions).

West of the Breggia outcrops is a broad valley, filled with Pleistocene deposits. The important Lugano fault runs north-south along this valley. On the western side of the valley, only a few kilometers from the Breggia gorge, the Jurassic is again exposed, but in a very different facies (Fig. 5, left-hand column). Instead of the lower Liassic cherty limestones which attain 1000 m or more, there are only thin lower Liassic and Pliensbachian deposits, filling karstic cavities in the deeply eroded Norian dolomites. (The very complex stratigraphy and lithology of these karst fillings is being studied in detail by F. Wiedenmayer of Basle.) This very sudden change of facies can hardly be explained otherwise than by assuming that the Lugano fault is a very ancient feature and was already active during Liassic time, with downthrow to the east. The western block was tilted westward, as the Liassic erosion is strongest just west of the fault.

The fault had almost no influence on the facies of the formations from the Domerian upward, except that thicknesses are somewhat greater in the downthrown block. This can mean either that the displacement diminished in the middle Lias, or that the whole area became deeply submerged. A submarine difference of level of, say, 300 m would change the facies of the sediments on either side if the sea were shallow, whereas at depths of, say, 2000 m such a change would presumably have little effect on sedimentation, or might at most give rise to some slumping.

Marine Breccia Formations

Characteristics and environment.—Schistes lustrés and leptogeosynclinal deep-sea deposits are not the only formations characteristic of the eugeosynclinal belt of the Alps, dating from the times preceding the early synorogenic Flysch phase. Polygenic marine breccias are perhaps just as significant for the understanding of the structural history of the geosyncline (Arbenz, 1919, p. 261).

These breccias are of varied origin. Intraformational breccias are common in many Triassic limestones and dolomites (Helvetic: Widmer, 1949; Briançonnais belt: Debelmas, 1952; 1955; Genge, 1958; Austroalpine nappes: Cadisch *et al.*, 1919; Kappeler, 1938; Sander, 1936). For most such breccias, break of thixotropy provides a satisfying explanation (Genge, 1958, p. 190); they can develop vertically and laterally out of normal strata and commonly also show evidence of slumping. There was apparently a considerable amount of seismic activity during the Triassic, culminating toward the end of the Ladinian. Other monogenic limestone-dolomite breccias may be due to diagenetic replacement (Gignoux and Avnimelech, 1937), and certain Triassic breccias are certainly terrestrial.

We are here, however, concerned with marine polygenic breccias of Jurassic and Cretaceous age. These again show very different characteristics. The matrix consists of shales, limestones, or even cherts; there is little detrital quartz, and the breccias are not associated with sandstones. Among the detrital components, Triassic carbonate rocks are often preponderant; basement rocks may play a

large part in some Cretaceous breccias. These elements are commonly very closely packed, and the matrix in most instances seems secondary. The rounding of the components is imperfect or nonexistent, the sorting poor to very good. Graded bedding is common. These marine breccias make up isolated beds or lenses, whole formations, or thick, poorly stratified and narrowly localized bodies.

Breccias of this type occur in almost every tectonic unit of the Alps, and they range in age from Rhaetic to Late Cretaceous. Nevertheless, there are certain zones in which Mesozoic breccias are very abundant and make up a large part of the rock sequence. Three belts are thus characterized:

(1) The outer margin of the complex Briançonnais platform in the broad sense—that is, the inner margin of the Valais trough: Oxfordian Brèche du Télégraphe in the Subbriançonnais of Savoy (Barbier, 1948; Schoeller, 1929), Upper Jurassic and Lower Cretaceous of the Falknis nappe in Graubünden (Allemann, 1956; D. Trümpy, 1916).

(2) The inner margin of the Briançonnais platform and outer margin of the Piemont trough: breccias of the "Quatrième Ecaille" near Briançon (Lemoine, 1951), Cretaceous Tsanteleina breccia of the Vanoise (Ellenberger, 1958, p. 327), Jurassic breccias of the Breccia nappe in the Prealps, discussed below.

(3) The border zone between the eugeosynclinal Piemont trough and the Austroalpine realm: Dolin breccias in the Dentblanche nappe of Valais (Hagen, 1948), Maran breccia of Arosa zone, and numerous breccias of different age in the lower Austroalpine nappes of the Upper Engadine.

Even in the comparatively tranquil, miogeosynclinal Helvetic realm, breccias, mainly of Early and Middle Jurassic age, occur in the vicinity of ancient fault scarps, and some of the microbreccias are turbidity-current deposits (Carozzi, 1957). The same is true for the upper Austroalpine nappes.

The localization of pre-orogenic marine breccias along the limits between platforms and deeper-lying troughs certainly provides a clue to their genesis. It is linked to the existence of tectonic scarps, in most instances contemporaneous normal faults and flexures. Some of the breccia bodies are simply submarine scree deposits, and they share the absence of medium-sized detritus with subaerial scree. Others may be due to shattering and mixing of consolidated or semiconsolidated rocks by earthquakes.

Many geologists have considered the breccias of the Alps as reliable indicators for shallow-water environment. They have interpreted these breccias as coastal-abrasion products, as deltaic deposits, or as having formed by wave action on shallow platforms (*e.g.*, Tercier, 1952). The association of breccias with radiolarian cherts was in turn held to be a convincing argument against deep-water origin of the latter. In fact, most Alpine breccias lack the characteristics of shallow-water psephites, notably the inevitable rounding of the components. R. Trümpy (1955a) and Allemann (1956, p. 108) have insisted on the notion that polygenic marine breccias could form at any depth, provided there was a sufficient slope and contemporaneous tectonic and seismic activity. Interbedding of breccias with radiolarian cherts, maiolica limestones, and shales without bottom fauna denotes deep-water environment for the former rather than shallow-water environment for the latter, especially where these breccias show coarse graded bedding.

Jurassic of the Breccia nappe.—The Breccia nappe is one of the higher units of the Prealps. It is preserved mainly in the Chablais arch south of Lake Geneva and in a smaller area in the Bernese Oberland south of Lake Thun. Very good accounts of its stratigraphy have been given by K. Arbenz (1947), Chessex (1959), Lugeon (1896), and Schroeder (1939). The exact provenience of the Breccia nappe is still unknown, but there is good circumstantial evidence for deriving it from the most westerly part of the Piemont eugeosyncline, adjoining the Briançonnais platform.

The post-Triassic formations of this remarkable structural unit are listed in Figure 6. Shallow-water sediments with benthonic fossils were formed up to the Sinemurian (early Lias). The "Schistes inférieurs" are probably a good example of nonmetamorphic Schistes lustrés (although without ophiolites). Crinoidal and bryozoan fragments in calcarenitic limestones are the only fossil remains.

The same kind of rocks constitute the layers between the breccia beds and the matrix of the breccias in the "Brèche inférieure." The lower limit of this formation is certainly diachronous. The breccia members consist mainly of angular fragments of Triassic dolomite and of Liassic limestone; only boulders derived from penecontemporaneous, poorly consolidated rocks show some rounding. The thickness of individual beds changes abruptly. Some coarse and irregular graded bedding is common but not general; medium-grained breccia passes upward

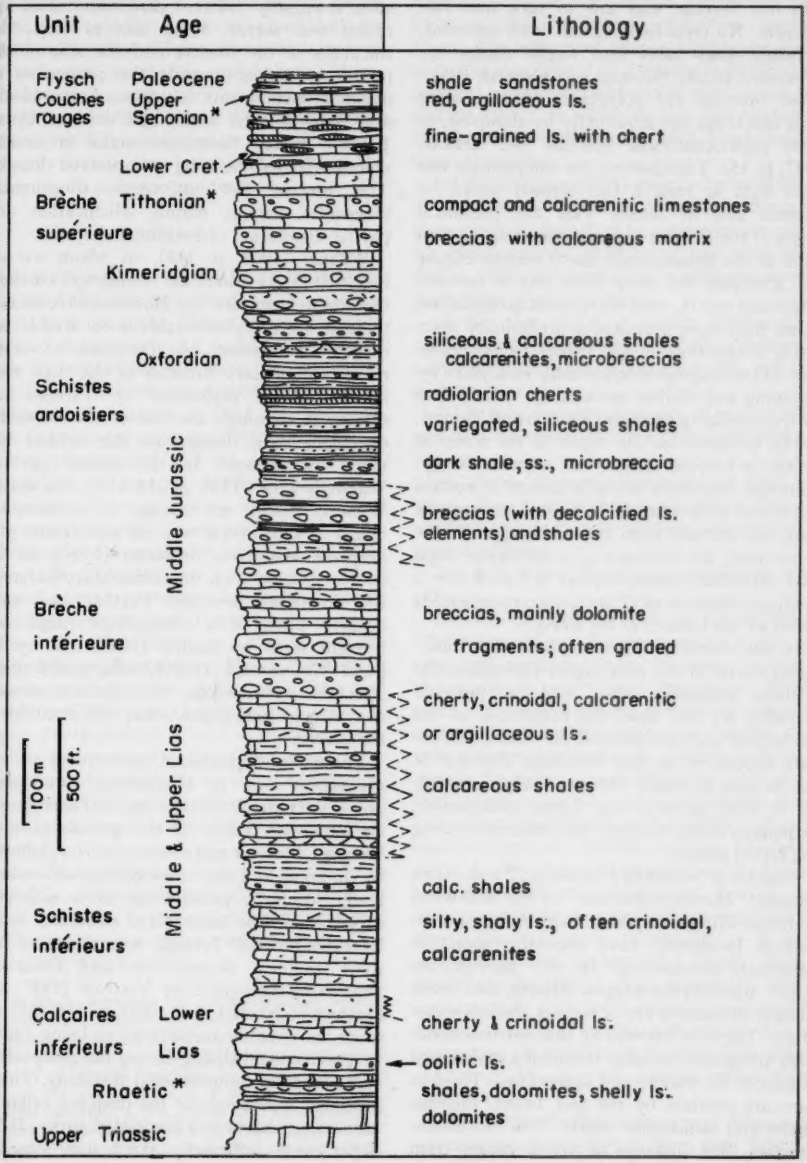


FIGURE 6.—JURASSIC AND CRETACEOUS SEQUENCE OF THE BRECCIA NAPPE IN ABONDANCE VALLEY (CHABLAIS)
Mainly after Chessex (1959)

into fine breccia, and this in turn into calcarenite. No cross-bedding has been recorded, although load casts and ripple marks are developed locally (Kuenen and Carozzi, 1953). Most breccias are polygenic, but limestone beds can break up, apparently by slumping, to form intraformational breccias (K. Arbenz, 1947, p. 15). Transport of the components was from west to east, a fact already noted by Lugeon and in accord with the presumed origin of the Breccia nappe on the steep eastern flank of the Briançonnais rise (*Compare* Figure 12). Probably this steep slope was of tectonic origin and was the seat of repeated earthquakes. These may have provoked a preliminary shattering of the Triassic and Liassic rocks. Transport of the fragments apparently took place by slumping and sliding movements, rather than by true turbidity currents (Kuenen and Carozzi, 1953). Evidence on the depth of the water in which the breccias came to rest is contradictory. Crinoidal fragments are so abundant in certain limestone beds that it is logical to suppose that the animals lived not far away. On the other hand, the existence of a submarine slope that permitted coarse detritus to travel over a minimum distance of 12 km implies considerable depth at the bottom of the slope.

In any event, the siliceous shales and radiolarian cherts of the next higher formation, the Schistes ardoisiers, may well be deep-sea deposits. By this time, the conditions on the sea bottom were unfavorable for calcium-carbonate deposition, so that limestone elements in the breccia members are commonly corroded. As in most parts of the Alpine geosyncline, carbonate sedimentation was resumed during the Late Jurassic.

Breccias of Sonnwend Mountains, Tyrol.—The Jurassic "Hornsteinbreccien" of the Sonnwend or Rofan Mountains above Lake Achen, northeast of Innsbruck, have been the object of passionate controversy. In this part of the upper Austroalpine nappes, Rhaetic and lower Liassic formations are of normal shallow-water origin. They are followed by thin red limestones with manganese nodules, containing cephalopod faunas of the middle and upper Lias. These in turn are overlain by red and brown siliceous shales and radiolarian cherts. The Hornsteinbreccias, the thickness of which ranges from 10 to 150 m, are interbedded with these cherty rocks. Above it are thick limestones with chert bands, of Late Jurassic age.

As a rule, the breccias are poorly bedded, except for their lower part, where they alternate with radiolarian cherts. The angular components, essentially different carbonate rocks, are rather well sorted. From east to west, the thickness of the breccia and the size of the fragments decrease, while the proportion of dolomite components increases. Large blocks and slabs of older rocks, and even of Upper Jurassic cherty limestones occur in several places. Graded bedding is observed locally. The breccia has undergone considerable diagenetic change, mainly silicification and partial dissolution of limestone elements.

Wähner (1903, p. 118), to whom we are indebted for an admirable monograph on these mountains, regarded the Hornsteinbreccia as a tectonic breccia, attributable to the Cretaceous orogenic movements. At the time he wrote, many sedimentary breccias in the Alps were interpreted as "mylonites" by Termier and others. A thorough account of the historical development of thought on this subject was given by Spengler in the second part of Wähner's work (1935, p. 118–153). One reason Wähner would not accept a sedimentary origin of the breccia was its association with deep-sea sediments. Ampferer (1908), on the contrary, insisted on the sedimentary nature of the Sonnwend breccias. Further and more decisive proof of a sedimentary origin was brought forth by Sander (1941) and by his pupil Weynschenk (1949),[8] who noted resedimentation phenomena, but who also stressed the strong post-depositional deformation of the breccias.

These post-depositional movements are not necessarily due to Cretaceous (pre-Gosau) orogeny. Heritsch (1915) suggested slumping as an important factor in the genesis of these breccias. Tectonic and seismic activity, followed by slumping and slow mass-gliding movements, really seems to provide the most satisfying explanation. The complicated structures in the leptogeosynclinal Jurassic formations of the more easterly Kammerkar and Osterhorn mountains discovered by Vortisch (1927, and abstract in Schaffer *et al.*, 1951, p. 354–357) are much more easily understood as being caused by gravitational sliding during the geosynclinal stage than by compressional thrusting. This is a further indication for the deep-sea origin of radiolarian cherts and associated rocks. If the slopes were sufficiently steep and long for large-scale gliding, and if the facies remained

[8] The rather curious criteria invoked by Weynschenk for proving shallow-water origin of all formations in the Sonnwend Mountains were severely criticized by Grunau (1959).

essentially uniform over wide areas, the depth of the sea must have been so great that the differences in level did not affect the nature of the sediments. In the Kammerkar and Osterhorn mountains such conditions obtained only during the Middle Jurassic; the Liassic formations show different facies on top of Rhaetic bioherms and in the depressions between the bioherms, 70 m lower, at most.

Flysch

General statement.—To the inhabitants of the Bernese Oberland and of Obwalden (southeast of Lake Lucerne), Flysch means a more or less shaly rock, producing moist acid soils and given to superficial slumping and flowing[8a]. This dialect term was introduced into geological publication by Studer in 1853 and has since been used in variable connotations (the historical development will be treated by W. Leupold in the Lexique stratigraphique international). There is now a kind of general agreement among Alpine geologists on the definition of the term, although this definition cannot be laid down in a few words. Some misuses persist, however; thus normal neritic or pelagic formations of Tertiary age are sometimes still called Flysch (*e.g.*, south-Helvetic Einsiedeln and Blattengrat "Flysch," consisting of pelagic marl shales and a thick limestone member with large Foraminifera), and of course many geologists have applied the term to formations outside the Alps which show no resemblance whatever to Alpine Flysch (*e.g.*, the Purbeckian of the Jura, brackish and subsaline carbonate rocks with intraformational breccias).

More than any other Alpine term, the facies name Flysch has been applied to formations of other fold belts. A good "universal" definition of Flysch has recently been advanced by Kuenen (1958, p. 329):

> "A thick sequence of pre-paroxysmal marine geosynclinal sediments, consisting of an alternation of evenly stratified shale and muddy sandstone (graywacke, etc.) and showing at least a moderate amount of graded bedding. The maximum grain size in the graded beds is 5 to 10 cm diameter. Coarser material is not graded and subordinate in amount. Transitions to or alternations with calcareous types also occur. Geological age is ignored."

In this paper, we are concerned only with the Flysch of the Alps. As Kuenen himself remarked, some Alpine Flysch formations fall short of the "universal" definition quoted above. The agreement among Alpine geologists as to the significance of the term is in part the result of an excellent and widely read review by Tercier (1947), in which he lists four essential characteristics of Flysch (p. 168). Freely translated, these are:

(1) Flysch corresponds essentially to detrital or terrigenous deposits, quite subordinately biogenic ones, deposited in a more or less regular alternation of sandstones and micaceous shales, with, sometimes, intercalations of conglomerates or of limestones.

(2) On the whole, it always represents a rather thick formation (group), the lithological units (formations) of which are poorly individualized.

(3) It is made up exclusively of marine sediments, in part neritic, but in part also of bathyal origin.

(4) From the paleogeographical point of view, Flysch must be considered a deposit proper to basins broken by steep and discontinuous cordilleras.

We shall follow Tercier by examining these points.

Lithological and biotal characteristics of Alpine Flysch successions.—Sandstones and shales, alternating regularly over, in some instances, thousands of meters, are the typical Flysch rocks. The sandstones are generally micaceous, commonly feldspathic, and some varieties contain a large proportion of detrital carbonate grains. Rounding is generally poor, except for the largest grains, and sorting is characteristically poor (Rech-Frollo, 1953; 1954). The cement is commonly calcitic, but may be argillaceous or silty. Most sandstones are associated with finer-grained detrital rocks, such as calcareous mudstones (commonly designated by the misleading term "Kieselkalk"). Typical graywackes are rather exceptional. Flysch shales are invariably silty, commonly sandy, and generally show mica flakes on the bedding planes. They may or may not contain a large proportion of calcium carbonate. The shale–sandstone ratio is very variable (from about 10:1 to 1:5).

The coarse detrital members of Flysch sequences generally show a very wide range of nature and size of components, among which basement rocks are commonly conspicuous. The elements are poorly rounded, and primary matrix is generally abundant.

Flysch limestones are generally of the "Alberese" type, thinly bedded and of exceedingly

[8a] Flysch never was a quarryman's term as suggested by Eardley and White (1947, p. 981).

fine grain. Genesis of these limestones is very enigmatic, particularly as they have as yet yielded no fossils, not even a pelagic microbiota.

Ophiolites occur only quite exceptionally in typical Flysch formations; they are characteristic of the older Bündnerschiefer group. Debris of intermediate to basic volcanic rocks, on the other hand, is very prominent in some Tertiary Flysch formations.

An important feature of Flysch sediments is the lack of an autochthonous, shell-bearing, bottom-dwelling fauna. Fossils of the Flysch belong to three categories:

(1) Tracks and burrows of bottom-dwelling animals, such as fucoids and helminthoids. Seilacher (1955) has shown that this type of organic trace indicates deep water, beyond the penetration of light.

(2) Shells of benthonic animals, mainly larger Foraminifera (*Siderolites*, *Nummulites*, etc.) and fragments of calcareous algae (*e.g. Lithothamnium*) and of bryozoans. These fossils occur in graded sandstones, and the mass of the remains (rather than their size) is of the same order as that of the sand grains (generally a few millimeters in diameter). Clearly, these neritic fossils were redeposited with the detrital matter, by turbidity currents and slumps. The scarcity of molluscan debris is noteworthy.

(3) Pelagic and characteristically microscopic fossils (Radiolaria, planktonic Foraminifera), some of which occur in graded sandstones and may be resedimented. Plant fragments are not rare in some Flysch formations and can be classed with category (1) (in coarse, basal layers of sandstones) or (2) (on bedding planes).

Formational and bedding characteristics of Flysch.—The thickness of Flysch sequences is generally considerable, greater than that of the earlier Schistes lustrés group. Flysch deposits of any tectonic unit measure between a few hundred and some thousands of meters, frequently 1000 to 2000 m. This great thickness ordinarily corresponds to a relatively short time span, covering a few stages (in rough figures, 5 to 20 million years). It is only where Flysch follows Bündnerschiefer without a break, as in the Prätigau, that a veritable "comprehensive series", representing, for instance, the whole Cretaceous epoch, is found. Flysch sedimentation is thus very rapid, by Alpine standards.

Flysch sequences are characterized not only by rapid, but also by more or less continuous, sedimentation. No major breaks are recorded by paleontological evidence. The different lithostratigraphic units are linked by passage beds, and the subdivision of the Flysch facies is almost as difficult as that of the Schistes lustrés (although less haphazard because of the fossils which diligent search will always reveal in the Flysch). No cyclothemic units of larger order can be observed; the normal pulsatory movements were obscured by the contemporary tectonic activity.

Owing to the tectonic complications and to the bad state of the outcrops, which are the bane of the geologist's life in Flysch country, it is difficult to follow a given unit or bed of a Flysch sequence very far. Nevertheless, at the scale of a quarry or even of a hillside, the individual beds show little change in composition and thickness. This characteristic contrasts strongly with the more limited lateral extent of most beds in the Molasse successions. The beds are rather thin—a few centimeters or decimeters—and fairly regular; sandstone members several meters thick are found only occasionally.

Graded bedding and the different sole markings caused by turbidity currents and bottom creep are common in most Flysch sequences (Allemann, 1956, p. 93–144; Crowell, 1955; Hsu, 1959; Kuenen, 1953; 1957a; Kuenen and Carozzi, 1953), but these features are not at all regularly developed, and the grading of Alpine Flysch sandstones is generally much less well developed than that of many Paleozoic graywackes. In many instances, only the topmost part of sandstone beds is graded. Erosional discontinuities at the base of sandstone layers are quite exceptional, and current ripple marks are rare; small-scale current-bedding is of some importance in the youngest Flysch formations. Evidence of slumping is not infrequent in the transitional facies to both Bündnerschiefer and Molasse; on the other hand, it seems to be absent from the more "flyschy," typical Flysch formations, in which the sandstones are in large proportion turbidites.

Bathymetric setting of Flysch sediments.—All Flysch sediments are of marine origin, and none shows evidence of abnormal salinity, except in zones of transition to Molasse, where the greatly narrowed Flysch sea became brackish. The bottom waters were generally well aerated, but in this respect, again, the youngest Flysch formations are abnormal and show lack of oxygen.

Up to the 1940's, Flysch formations were generally regarded as neritic shallow-water deposits. Tercier (1947, p. 179) was among the first to oppose these views, with very sound arguments. In his opinion Flysch sedimentation was not restricted to any particular depth zone but could take place along steep slopes, from

coastal environments down to great depth. With the extremely fruitful notion of turbidity currents taking its triumphal course, the pendulum has swung to the opposite extreme, and practically all graded sandstones of Flysch type are being considered as deep-sea sediments. This also is probably exaggerated.

As a general rule, Flysch sediments can be said to have formed below the reach of wave action. One important (although negative) piece of bathymetric evidence is the lack of an autochthonous shell-bearing fauna. In addition, turbidity-current deposits of great horizontal extent presuppose the existence of long slopes and imply great depths at the far end of these slopes. However, the abundance of resedimented larger Foraminifera and calcareous algae in some Flysch sandstones, and of well-rounded pebbles (among a majority of poorly rounded ones) in some Flysch conglomerates indicates that shallow-sea, perhaps even coastal, environments were not far from those where these Flysch sediments were laid down. The abundance of floated vegetal remains may also point in this direction, although it is by no means decisive. The author knows of no conclusive evidence of lime leaching in Flysch deposits, whereas such evidence abounds in the older, leptogeosynclinal formations; this, however, may be due to the accelerated rate of deposition. The author estimates that most Flysch rocks have formed at depths exceeding 200 m, but not more than a few kilometers. Flysch sedimentation marks the filling of the geosyncline.

Some shallow-water sediments do occur locally in Flysch formations. The Cretaceous Simme Flysch contains beds of polygenic conglomerates (the Mocausa conglomerates), with well-rounded pebbles, and layers with a rich bottom fauna of Cenomanian age (Campana, 1943; Jeannet, 1912–1913). These deposits might even be regarded as approaching the Molasse type of facies, related, of course, not to the principal, Tertiary folding phases of the western Alps, but to the mid-Cretaceous orogeny of the Austroalpine nappes, to which the Simme nappe perhaps should be attributed. Also, the northern Penninic or Ultrahelvetic Sardona flysch of the Glarus Alps—not a very typical Flysch in some respects—has one member, the Paleocene Sardona quartzite, consisting of clean, occasionally cross-bedded quartz sandstones (Leupold, 1937; 1942; Rüefli, 1959).

Tectonic setting of Flysch sediments.—Compared with the older miogeosynclinal and eugeosynclinal formations of the Alps, the Flysch sediments show the influence of contemporaneous tectonic movements much more clearly. The marine breccias of Jurassic and Early Cretaceous age discussed earlier also related to tectonic activity, but these breccias were restricted to certain belts in the vicinity of ancient fault scarps. During the time of Flysch sedimentation (Cenomanian to early Oligocene) the submarine relief was much stronger, and tectonic movements affected the whole of the Alpine belt, although not all of its parts at the same time. The swells and platforms of pre-Cenomanian time were submarine, or at best rose above the sea as low islands. When the Flysch was formed, large tracts of the original basin were emerged and undergoing active elevation and erosion. Flysch sedimentation immediately precedes in time the major structural revolution of each paleogeographic unit.

A very puzzling problem concerns the source of the sand in the Flysch sediments. Rising cordilleras would of course furnish a great amount of detrital matter, but it is hard to explain why and how most of this was reduced to comparatively small grain size before being resedimented by turbidity currents and slower mass-redepositional processes. Possibly much of the fragmentation of rocks took place in shallow waters surrounding the emerged island belts, and some of it by river transport. Some authors (especially Brückner, 1952) have considered a deltaic origin for Flysch materials. This idea was recently taken up by Kuenen (1957b; 1958), who has given much thought to the problem in connection with the proved longitudinal filling of many Flysch troughs by turbidity currents. Reconstruction of the paleogeographical configuration during "Flysch times," however, leaves little or no space for the existence of larger river systems inside or outside the present Alps. In many instances (*e.g.*, Niesen Flysch, Prätigau Flysch, part of Gurnigel Flysch), it can be shown that the detrital matter is derived laterally from the cordilleras, even if it may have been redistributed along the axis of the basins by turbidity (or other) currents. Much of this material was furnished by granites and other rocks of the pre-Triassic basement.

Relationship of Flysch to other facies groups.—No gradation of Flysch into normal, miogeosynclinal or platform deposits is known. Where the two types of formations enter into stratigraphical contact, the Flysch rests unconformably on the older rocks, down to the pre-Triassic basement.

We have already seen, however, that the Bündnerschiefer group of sediments may pass

upward and laterally into Flysch, and that it is quite impossible to draw a sharp boundary between the two.

The relationships of Flysch to pelagic formations, such as radiolarian cherts, maiolica limestones, and especially calcareous shales or argillaceous limestones with planktonic Foraminifera, are of prime importance. Sediments of this type in many places underlie Flysch formations. The boundaries are commonly sharp, or there may be lateral transitions and gradation by interbedding. Much of what is called "Flysch calcaire" in the western Alps consists of shale–limestone alternations, which may represent a transitional facies between pelagic sediments or Bündnerschiefer to typical Flysch. Where Flysch sedimentation took place over formerly miogeosynclinal or nongeosynclinal belts, there are pelagic sediments between the neritic formations and the Flysch (*e.g.*, Eocene Stad shales in the Helvetic Alps, Cretaceous to Paleocene Couches Rouges in the Prealps). This implies that the downbuckling of a particular belt preceded its filling by Flysch sediments.

Just as Bündnerschiefer pass upward and outward (north or west) into "pre-Flysch" and Flysch, the Flysch itself grades upward and outward into Molasse. The subsiding belt thus shifted toward the external border of the Alps and was more and more filled in by detrital matter from the rising chain.

There is nothing like "The Flysch" of the Alps, but a great many different kinds of Flysch. Their litbological and bedding features vary widely, their age from Cenomanian to early Oligocene. (Sedimentation of "pre-Flysch" type set in even earlier, during the Early Cretaceous.) Some of these formations are quite unequivocally Flysch; others represent particular types, and still others are transitional either to Bündnerschiefer (including "pre-Flysch") or to Molasse. Four examples illustrate these differences.

Niesen Flysch (transitional to "pre-Flysch").—The Niesen nappe is known only from the Prealps between the Rhone and Lake Thun, over a distance of 60 km along the strike. It lies between the Ultrahelvetic nappes below and the Médianes nappe above (Pl. 2), but its exact origin is still disputed. However, it can safely be assumed that the Niesen nappe is derived from the lower Penninic Valais facies belt, probably from the more external part of it. The most important monographs are those by Andrau (1929), Bornhauser (1928), McConnell (1951), and de Raaf (1934); recent research by Lombard (1942; 1949; 1956, p. 638) is still unpublished, save for a few preliminary notes.

The stratigraphic sequence of the nappe begins with isolated slices of crystalline basement rocks and a very unimpressive development of Triassic. It is followed by Liassic crinoidal limestones and nonfossiliferous, slightly metamorphic argillaceous slates, apparently of Jurassic and perhaps Early Cretaceous age. In some sections, these slates grade upward into the Niesen Flysch proper, which in other places rests unconformably upon the older formations. The Flysch itself ranges up to 2000 m thick. It differs widely in composition from the northeastern to the southwestern end of the chain, and correlation between the two areas has yet to be established. Most of it is of Late Cretaceous (especially Maestrichtian) age; the topmost part reaches into the Paleocene or early Eocene. Cenomanian *Orbitolina* has also been found (Lombard, 1942), but it is still difficult to judge how much of the group is older than Campanian.

Studer, in 1853, did not consider the "Niesen breccia" as Flysch, and the Niesen Flysch differs in many respects from other, more typical Flysch sequences. It forms a chain of bold, pyramid-shaped mountains (Pl. 2), contrasting with the soft hills of other Flysch areas. (The Eocene Aiguilles d'Arves Flysch of Savoy and the Cretaceous to Eocene Arblatsch Flysch of Graubünden give the same morphological effect as the Niesen Flysch.) Limestones form a large proportion of the Niesen sequence, and most of them are not of the sublithographic Alberese type; one unit consists essentially of rather thick-bedded limestones with calcified sponge spicules. Other benthonic shelly fossils, such as crinoid fragments and even brachiopods, are found at some places. Conglomerates are very prominent, especially at the southwestern end of the chain. Many of them contain enormous boulders of granite and of crystalline schists. Many of these conglomerates and breccias are reminiscent of the older scarp breccias discussed in a preceding section on the Breccia nappe. The coarse material is derived from the west—that is, from the external margin of the basin. About 20 km farther east, the lower part of the Niesen Flysch is formed by monotonous calcareous shales with a few beds of graded sandstone, but no more conglomerates. Grading in general is poor; on the other hand, small-scale cyclothems a few decimeters or centimeters thick have been described from the upper part of the Niesen Flysch (Lombard, 1949). When

complete, they show the sequence sandstone–shale–limestone, with many variations. Sole markings and load casts are abundant; contorted bedding is not uncommon. There is good evidence of breaking up of beds by penecontemporaneous slumping (*e.g.*, McConnell, 1951, p. 40).

Turbidity currents, while certainly active, do not appear to have been the essential agent of transport for the detrital matter in the Niesen Flysch, except possibly its uppermost parts. For *this particular type* of Flysch (or "pre-Flysch"?) bottom creep, as put forward by Lombard in his monumental work (1956), must certainly be considered as an important factor. The persistence of calcareous deposition through much of the group implies that depth of the sea was not excessive, at any rate not in the western part of the present chain. Lateral variations of facies are more abrupt than in the later and more typical "turbidite Flysch" formations.

Other Cretaceous Flysch and "pre-Flysch" formations of the Valais facies belt show rather similar characteristics: for instance, the middle part of the Prätigau schists (discussed earlier), or the Tarentaise breccias of Savoy, Aosta (northern Piemont), and Valais (Schoeller, 1929; R. Trümpy, 1954), which contains a unit of giant conglomerates having rounded boulders up to 35 m across.

Schlieren and Gurnigel Flysch sequences of central Switzerland.—The Schlieren and Gurnigel sandstone Flysch complexes can serve as examples for typical Flysch formations.

The Schlieren Flysch (Fig. 7) fills a large syncline on the Helvetic nappes between Lake Thun and Lake Lucerne; probable equivalents are also found south and east of Lake Lucerne. It rests upon wildflysch of late Eocene age and is in turn overlain by outliers of the Médianes nappe in the Giswiler Stöeke (Vonderschmitt, 1923). The Schlieren Flysch, 1500 m thick, thus constitutes a nappe of its own, and its original substratum is unknown. It may belong either to the internal part of the Ultrahelvetic or to the external part of the Penninic realm. Because of its rather simple tectonic structure, some good outcrops, and especially thanks to a monograph by Schaub (1951), its stratigraphic succession is better known than that of most other Flysch sequences.

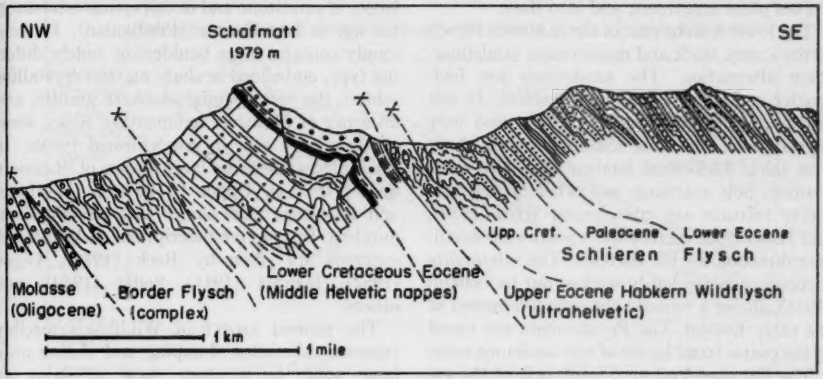


FIGURE 7.—CROSS SECTION THROUGH THE BORDER ZONE OF THE ALPS WEST OF LAKE LUCERNE
After Mollet (1921) and Schaub (1951). X indicates thrust fault.

The record begins with a Flysch of Late Cretaceous age, consisting of shales with helminthoids, sandstones of generally fine grain, and siltstones. Peculiar, very fine-grained quartzitic sandstones showing an oily luster on fracture planes ("Oelquarzit") are found in the basal part of the Schlieren group, and also in many Cretaceous Flysch formations of eastern Switzerland and of Austria. Their origin is very enigmatic. Sole markings are fairly well developed; some contorted lamination has been observed, but grading is generally poor and commonly absent. South of Lake Lucerne, coarse breccias with elements of basement rocks form an important part of this basal unit of the Schlieren Flysch (Geiger, 1956).

The Paleocene is represented by sandstones, commonly arkosic and very well cemented, in

thick beds (up to 10 m), with thin shale partitions. Many of the sandstone beds begin with microbreccias. The main body of the layer is characteristically of uniform grain size, but at the top there is regular grading into finer-grained, laminated sandstone, commonly with drifted plant fragments, and into shale.

The lower Eocene part of the Schlieren Flysch forms a very thick and monotonous sandstone–shale alternation. The sandstones are feldspathic and only moderately calcareous. In this portion, graded bedding is constant and very regular; each sandstone–shale sequence is about 2 m thick. Individual laminae may also show grading. Sole markings and other typical turbidite features are conspicuous (Hsu, 1959), and Hsu (1960) established a southwest–northeast direction for the currents. The nummulite succession, meticulously worked out by Schaub (1951), shows a remarkably complete record of the early Eocene. The Foraminifera are found in the coarse basal layers of the sandstone beds.

The Gurnigel-Voirons Flysch belt of the external Prealps (Gagnebin, 1924; Lombard, 1940; Tercier, 1928) consists mainly of a sandstone Flysch very similar to the Schlieren Flysch. There is no Cretaceous Flysch, but on the other hand the succession reaches up into the middle Eocene. As a rule, the polygenic sandstones are more calcareous and coarser than those in the Schlieren Flysch. Graded bedding is irregularly developed, but sole markings are common. Direction of turbidity currents is mainly northwest-southeast (Crowell, 1955; Hsu, 1960). This led Hsu to assume that the Gurnigel and Schlieren Flysch basins lay on different sides of the same cordillera, on which, among other rocks, the Habkern granites cropped out. The Gurnigel Flysch would thus become of more internal origin than the Schlieren Flysch. For structural and paleogeographical reasons, the present author favors another hypothesis: Gurnigel and Schlieren Flysch were laid down in the same basin, the former to the northwest and nearer the "Habkern granite islands," the latter farther southeast and east, as today, and presumably in deeper water. The different transport directions might be attributed to transverse turbidity currents on the northwestern slope of the basin, changing to the longitudinal southwest-northeast direction in the basin axis.

Wildflysch of central Switzerland.—The type Wildflysch (*vildfleesh*) belongs to the Ultrahelvetic nappes of central and western Switzerland. It is generally overlain by the Schlieren and Gurnigel Flysch sequences; the former constituting certainly, and the latter probably, a higher nappe (Fig. 7). The term was introduced by Kaufmann in 1870 (Cadisch, 1953, p. 174). "Wild" refers to the bedding, which is extremely irregular or chaotic. Wildflysch consists mainly of shales, generally poor in carbonate mineral content, with torn and contorted beds and lenses of sandstone and of calcareous mudstone. Its age is late Eocene (Priabonian). It commonly contains large boulders of widely differing type, embedded in shaly matrix: crystalline schists, the very peculiar Habkern granite, and an array of Mesozoic sedimentary rocks, some of them very rich in nondeformed fossils. In other places, there are large masses of "Leimern beds", that is, Upper Cretaceous limestones and Paleocene marl shales with pelagic Foraminifera. Instructive descriptions of Wildflysch outcrops are given by Beck (1911), Gigon (1952), Jeannet (1941), Soder (1949), and others.

The general aspect of Wildflysch forcibly suggests submarine slumping and sliding on a large scale, on a steep slope adjoining an actively rising ridge. It is not yet certain whether this slope faced north or south. One measured granite boulder is 13,000 m³, and some of the slide masses of Leimern beds are more than 100 m long. The sliding evidently took place during the sedimentation of the Wildflysch; Buxtorf (1943) has described a beautiful transgression contact between Leimern limestones and Wildflysch. These contemporaneous block slides are reminiscent of the "olistrostromi" of the Apennines. The main thrusting and gravity sliding of the Ultrahelvetic nappes occurred soon after the deposition of the Wildflysch.

Wildflysch facies is not limited to the upper Eocene of the Ultrahelvetic nappes, but it is not displayed elsewhere so characteristically. Normal Flysch rich in shales can assume a Wildflysch-like habit when subjected to slumping or to strong and irregular tectonic deformation.

Flysch of the northern Helvetic belt and passage to Molasse.—These Flysch masses (Fig. 8), which attain a thickness of nearly 2000 m, are the youngest and originally the most external (northerly or westerly) of the Alps. Their age is latest Eocene to early Oligocene. By this time, the main body of the future chain already was above sea level and had undergone considerable folding and thrusting. The axis of subsidence had shifted northward to the site of the present Aar, Montblanc, and Aiguilles Rouges massifs. The foredeep continued to wander outward during sedimentation of this Flysch, and it reached the inner margin of the Molasse basin in the middle Oligocene.

The Tertiary formations in this region are markedly heterochronous (Fig. 8). They begin with normal neritic deposits: limestones and glauconitic quartz sandstones with large Foraminifera (*Nummulites*, *Assilina*, *Discocyclina*) and calcareous algae. These are lower Eocene in the originally southeastern part of the Helvetic realm, middle Eocene in a median belt, and lower upper Eocene in the northwest; lines of invariant facies (isopic lines) cross the structural trend at angles up to 35°. These neritic formations are overlain by pelagic marl-shales with *Globigerina* (Stad shales), equally heterochronous, which in turn pass abruptly upward and southeastward into sandstone Flysch.

This autochthonous[9] and north Helvetic Flysch is poor in carbonate minerals. Its lower formation, the Taveyannaz sandstones, consists of thick-bedded, greenish, mottled sandstones with partings of argillaceous and silty shale. Shale fragments, embedded in a semi-consolidated state, are common in the basal layers of sandstone. Graded bedding and sole markings are not conspicuous. Many sandstones of this and the next higher formation are graywackes, a type of rock elsewhere rare in the post-Paleozoic series of the Alps. The Taveyannaz sandstones are remarkable for being composed mainly of detritus of andesitic and spilitic lavas, among others pillow lavas (de Quervain, 1928; Vuagnat, 1952). Many of the spilitic lavas are strikingly similar to nonmetamorphic ophiolites from the upper Penninic nappes. The Taveyannaz sandstones pass upward into the Altdorf or Val d'Illiez sandstones, where the volcanic material is gradually replaced by more polygenic detritus, among which fragments of radiolarian chert are noteworthy. Silty shales without carbonate are more abundant here and commonly show regu-

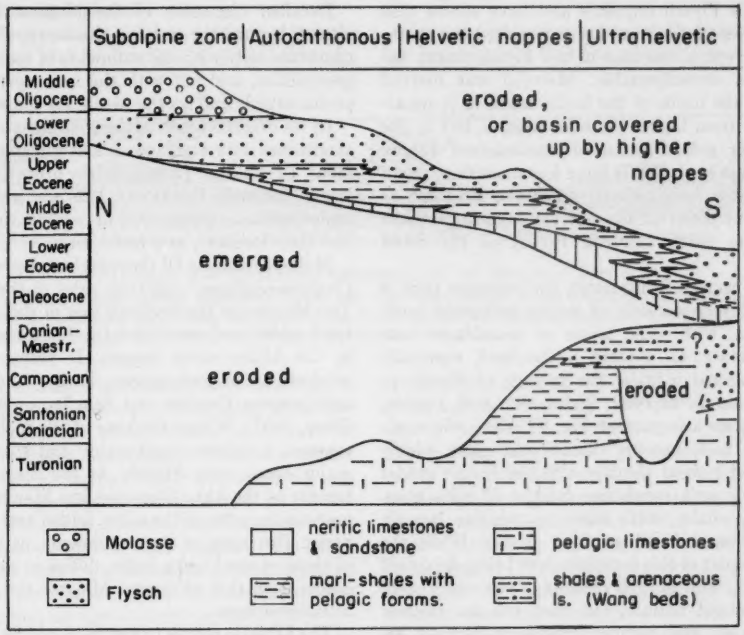


FIGURE 8.—AGE RELATIONSHIP OF FLYSCH FORMATIONS IN THE HELVETIC BELT OF EASTERN SWITZERLAND

The horizontal distance is about 70 km. After Brückner (1952), much modified (based in part on results of Leupold and his students).

[9] In eastern Switzerland, tectonic structures of the Flysch are quite disharmonic with regard to those of the Mesozoic cover of the Aar massif, and the Flysch cannot be considered as strictly autochthonous (Brückner, 1943; Styger, thesis in press).

lar graded beds of millimeter thickness. The sandstone beds are characteristically graded, and there is also much slump folding with truncated tops and small-scale current-bedding. Brückner (1946), Moret (1934), and Vuagnat (1952) have worked out the lithostratigraphy of this Flysch sequence and have shown that the older formations lap out toward the northwest, where the base of the Flysch finally becomes unconformable. Material was derived from the inside of the basin, much of it apparently from higher Penninic nappes, but in the similar grès d'Annot of southeastern France Kuenen *et al.* (1957) have found south to north —that is, longitudinal—transport. Research on current direction has not yet been undertaken in the north Helvetic Flysch of the Swiss Alps.

In eastern Switzerland, the youngest part of the Flysch consists of coarse polygenic sandstones, with some layers of roundstone conglomerate. In western Switzerland, especially in the Val d'Illiez, the passage of Flysch to Molasse is exposed (Schroeder and Ducloz, 1955; *see* also present Fig. 8). Here, the sandstone beds become thinner and more widely spaced toward the top, and the Flysch grades into a unit consisting mainly of micaceous, marly shales, with fishes, euryhaline lamellibranchs, and drifted land plants. While the lower part of this formation was being deposited salinity and oxygen content of the water were still about normal, but then the sea became brackish, and bottom conditions became reducing. These bituminous shales are overlain by marine sandstones in regular beds, without grading, above which begins the terrestrial, middle Stampian formation of the Molasse rouge.

The micaceous shales and overlying sandstones represent the very last deposits of the Alpine geosynclinal sea. All that is left of its ancient glory is a narrow strait of brackish and shallow water, and this last remnant is finally filled by the detritus from the rising Alps. But the same formation of Rupelian (lower middle Oligocene) shales and sandstones that was laid down in this expiring Flysch Sea appears in many places in Switzerland and Bavaria along the inner margin of the Molasse basin, where it is called "Lower Marine Molasse." The marine sandstones in its upper part are very consistent, in spite of changing names such as grès des Carrières, grès de Vaulruz, Horwer Platten, and Bausteinzone. Nothing could better illustrate the facies development of this formation, intermediate between Flysch and Molasse, than this equivocal nomenclature; the same beds are called Flysch in the Val d'Illiez and Molasse elsewhere.

Molasse

Detailed discussion of the Molasse formations is beyond the scope of this work, which is concerned only with the sediments of the Alpine geosyncline, and not with the synorogenic deposits outside and on the mountain chain.

In its original sense Molasse designated soft sandstones with argillaceous and calcareous cement. In southern France, on the other hand, it is used for shelly limestones. It is now generally understood as a group term or, outside the Alps and their foreland, as a facies term.

Molasse deposits fill the marginal depressions ("exogeosynclines") on both sides of the Alps. The Molasse at the northern foot of the Alps is itself folded and overridden by the frontal part of the Alpine cover nappes. It reappears in windows under these nappes, in the Val d'Illiez and between Gresten and Scheibs, in Austria (Prey, 1957). Where the base of the molasse is exposed, it follows conformably and without a major break upon Flysch. At the eastern extremity of the Alps, Oligocene and Miocene formations lap over on the older, folded and thrust rocks. The facies of these sediments, as well as of those of the Vienna basin, differs in some respects from that of typical Molasse and is not considered here.

The Molasse ranges in age from middle Oligocene to latest Miocene. Near the border of the Alps, thicknesses are enormous—4500 m for the middle and upper Oligocene in eastern Switzerland (Habicht, 1945). Sedimentation and "foot-by-foot" subsidence were more rapid than in any part of the geosyncline proper. Away from the Alps, the Molasse thins abruptly and measures only 1500–2000 m in the central part of the Molasse basin.

Only the most important differences between Flysch and Molasse are mentioned here. It is understood that the two group or facies types are linked by passage beds, but nobody will confuse a typical Flysch and a typical Molasse outcrop.

(1) The Flysch sediments formed within the Alpine geosyncline; Molasse belongs to the marginal depressions outside the chain or (in the eastern Alps) to basins on the Alps themselves.

(2) In consequence of this, Flysch is entirely marine, Molasse mainly continental. The marine Molasse formations were laid down in

shallow, temporarily brackish seas, except for regions like the Vienna basin.

(3) In a given area, Flysch sedimentation begins with the first folding movements and ends before the most important phases of tectonic deformation. Molasse was formed during and after these phases.

(4) Except for the different degree of induration some Molasse sandstones can be strikingly similar to some Flysch sandstones. Argillaceous and silty sandstones as well as sandy shales are common in the Molasse; coal beds, some of them workable, are frequent; limestones (shelly and fresh-water limestones) are negligible.

(5) Conglomerates with well-rounded pebbles (Nagelfluhen) are abundant and make up whole mountains along the inner margin of the basin. Most of them are fluviatile, deltaic deposits; others are marine shore-zone conglomerates. Both types are absent from typical Flysch.

(6) Graded bedding, so typical of many Flysch formations, has not been recorded from the Molasse. Cyclothems of the "Pennsylvanian" type, starting with an erosion surface and, if complete, having a coal or fresh-water limestone member in their upper part, are common (Bersier, 1945; 1949; 1950). Washouts, coarse and irregular cross-bedding, and ripple marks (both current and wave) are well developed.

(7) The bedding of the Molasse formations is generally thicker than that typical of the Flysch, and individual sandstone, conglomerate, or shale members may reach 50 m. On the other hand, the lateral extent of beds is very limited except for a few marker beds like fresh-water limestones or rare volcanic ash beds.

For all problems bearing on the sedimentation, stratigraphy, and paleogeography of the Molasse group, the reader is referred to the numerous papers on these subjects, from which only a few recent works of general importance are here cited: Abele *et al.*, 1955; Büchi, 1950; Fuechtbauer, 1954; Hofmann, 1957; Lemcke *et al.*, 1953; Rumeau, 1954; Rutsch, 1946; Speck, 1953.

Structural Evolution during the Geosynclinal Phase

Argand's Theory of Embryotectonics

Since the end of the nineteenth century, it has become clear that the original Alpine geosynclinal belt included a number of longitudinal swells and basins. Haug (1900; 1911) was the first to establish a succession of (partial or intra-) geosynclines and "geanticlines." (This use of "geanticline" for linear welts inside a geosynclinal belt, of course, does not conform to Dana's original definition.) Haug considered these elements as more or less symmetrical. Emile Argand (1916; 1920) proposed a coherent theory on the origin and the structural significance of these paleotectonic units (Fig. 9). In his view, the "geanticlines" were nothing more than incipient ("embryonic") anticlinal folds on the bottom of the geosyncline. These folds are held to be asymmetric, with a steep external flank and a more gentle slope toward the south or east. To him they represented the early stages of development of the great nappes (essentially the Grand St. Bernard and Dentblanche nappes). "Neritic" formations are found along the front (exterior) of these nappes, "bathyal" ones on the back (interior) and in the synclines between nappes. From the first faint swells of Pennsylvanian time to the last tectonic convulsions in the Miocene or Pliocene Argand saw a perfectly straightforward line of events in which asymmetric folds became exaggerated into recumbent folds and thrust surfaces developed gradually along the inverted limbs. (Argand considered all nappes as great recumbent folds, a view no longer accepted.) Ophiolite magma followed these thrust surfaces and flowed out in the deepest internal part of the individual troughs. During the Tertiary paroxysms, these thrust folds continued to move at an

Figure 9.—"Embryonic Geanticlines" in Mesozoic Time

Copied from Argand (1916). Northwest to the left. The "geanticline" in the center of the diagram represents the Briançonnais belt, along the front of the future Grand St. Bernard nappe; the one to the right prefigures the front of the Dentblanche nappe. Ophiolites in black.

accelerated rate and finally became the great Penninic nappes with basement cores. Some of these nappes—the Simplon nappes and the Monte Rosa nappes—came into being after the others and had no embryonic prefiguration in Mesozoic times.

Argand's theory certainly was a stroke of genius and quickly won wide acceptance. It furnished a beautiful means of explaining Alpine paleogeography, imputing all the vicissitudes of Alpine history to one process and to one only, the "impact of Africa on Europe." It was enthusiastically taken up by Staub (1917), who applied it to Graubünden and who recognized a far greater number of individual geanticlines and geosynclines than Argand had, each of the swells being now represented in the frontal part of a nappe. Embryotectonics came near to being accepted as an established fact and played a large part in the brilliant syntheses by Gignoux (1950), Moret (1954), Staub (1937; 1942), and Tercier (1952).

In the success of this theory, in spite of Haug's judicious warnings (1925, p. 243), it was often forgotten that it had been founded on very scant and controversial stratigraphic evidence. Detailed research in different parts of the Alps failed to prove the existence of embryonic folds, such as Argand had visualized. Many submarine or temporarily emergent welts were discovered, but they were not systematically asymmetric, did not show anticlinal structure, were not all persistent, and are not necessarily linked to the fronts of nappes. Many steep scarps were shown to face the wrong way —toward the south or, in the western Alps, toward the east. In the meanwhile, the study of other mountain chains, presenting a more easily readable paleogeographic record because of less extreme tectonic complications, produced no evidence of crustal shortening during the earlier phases of geosynclinal history (Bucher, 1933).

In the Alps, two factors were instrumental in the formulation of a new concept of paleotectonics: the discovery of Mesozoic normal faults in the Helvetic realm by Günzler-Seiffert (1941; 1952), and Lemoine's (1953) scale reconstruction of the Briançonnais "geanticline" (1953), which showed a broad, gently undulating submarine swell limited by scarps on either side. Almost simultaneously Debelmas (1957), Schoenenberg (1958), and R. Trümpy (1955a; 1957; 1958) concluded that the theory of embryonic prefiguration of Alpine nappes should be abandoned, at least as a dogma.

The criticism of the theory of embryotectonics refers to the paleogeographical configuration of late Triassic, Jurassic, and early Cretaceous time. In Argand's original concept, crustal shortening by tangential compression had already set in during the late Paleozoic. Later authors have postponed the onset of embryonic folding to the Triassic or to the early Jurassic. But in the new interpretation of Alpine pre-orogenic history, the structural pattern of the geosyncline in the Jurassic and Cretaceous is ascribed essentially to radial or tensional stresses. On the other hand, there is ample evidence of genuine folding during the later geosynclinal phase, when the Flysch was laid down (Cenomanian to early Oligocene). In discussing the structural evolution of the geosyncline, we must distinguish between these two phases.

Importance of Mesozoic Faulting

Mesozoic faults in the Helvetic realm.—Günzler-Seiffert (1941; 1952) was the first to draw attention to the importance of Mesozoic faulting in the Alps. His studies of the Helvetic nappes of the Bernese Oberland led him to recognize three major longitudinal east-northeast to west-southwest breaks affecting thickness and facies of the Jurassic and Cretaceous formations (Fig. 10). These faults were reactivated during the Tertiary orogeny when they had a strong influence on the fold structures. Downthrow of the faults was originally to the south, and the individual fault blocks were tilted northward. South of each fault, the formations are much thicker and more nearly complete than to the north. According to Günzler the movements along the faults were synchronous, discontinuous, and coincide with Stille's (1924) orogenic phases: early Cimmeric phase proved by the localized development of Upper Bajocian, Bathonian, and Callovian;[10] late Cimmeric phase demonstrated by lack of Tithonian; Austric phase established by lack of middle Cretaceous, etc. The tensional effect of the fault movements is clear enough for Jurassic time, but the later and weaker Cretaceous and Eocene movements are not so easy to interpret.

Mesozoic faulting on a large scale has also been reported by Baer (1959) from the western end of the Aar massif.

In the Helvetic nappes of the Glarus Alps, in

[10] Stille's early Cimmeric phase is supposed to have taken place between the Triassic and the Jurassic. This is just one example, among many, showing how far the phase concept will be stretched to fit the observations of its believers.

eastern Switzerland, pre-Alpine faults have been described independently from Günzler's pioneer work. Stratigraphic knowledge of this area is still much indebted to the masterly works by Arnold Heim (1916) and Oberholzer (1933). In Pennsylvanian and Permian time, part of the present Glarus Alps was occupied by a late Hercynian fault trench, which was filled by the thick continental formations of the Verrucano. Movements along the north-northeast to south-southwest trending faults which bordered this trench ceased before the Triassic, which is represented by thin terrestrial and shallow-water marine deposits with evaporites in the southeastern part. Liassic formations are absent from the cover of the autochthonous Aar massif and from the lower (originally more northerly) Helvetic nappes. In the higher nappes is 450 m of Liassic sandstones, shales, and arenaceous limestones, all shallow-water deposits. The Liassic formations appear very suddenly (R. Trümpy, 1949). The Lias-free northern area was probably part of a large, low island extending across much of the future Molasse basin. To the southeast, all Liassic formations appear almost along the same line and thicken abruptly. Maximum thickness is reached only a few kilometers from the margin of the subsiding basin. Breccias with elements of Triassic and (rarely) Permian rocks also occur along the hinge line between the stable, probably emerged platform in the north and the miogeosynclien in the south. This flexure scarp was active during the Early Jurassic, always with downthrow to the south; movements were continuous, although not of uniform intensity. Farther south, a pattern of antithetical flexure steps denote tensional deformation.

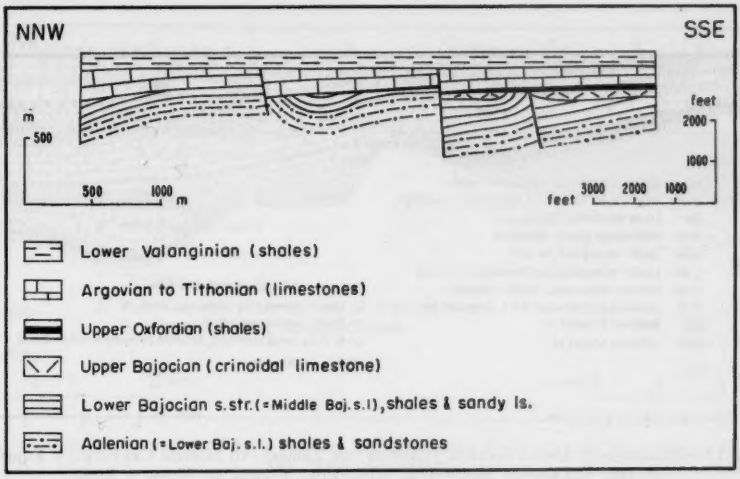


FIGURE 10.—PALINSPASTIC SECTION THROUGH THE MIDDLE AND UPPER JURASSIC FORMATIONS IN THE WILDHORN NAPPE OF THE BERNESE OBERLAND, SHOWING THE MESOZOIC FAULTS
After Günzler-Seiffert (1941)

The Aalenian transgression covered all the region of the Glarus nappes (indeed, almost the whole of the Helvetic realm) with its black, pyritic shales. Some of the Liassic flexures do not appear to have become reactivated. New faults came into existence in the late Aalenian. Detailed mapping and stratigraphic research by Schindler (1959) in the Glärnisch Mountains has given remarkable data on these Mesozoic faults (Fig. 11). One fault had a downthrow of more than 100 m in the Middle Jurassic; south of the fault, the Bajocian is almost ten times thicker than it is immediately north of it. Some breccias are found in the downset limb, and Schindler (1959, p. 22) also records slumping along the fault scarp, but this second point is not quite certain. The fault angle was rather low (20–45°). The fault blocks are tilted to the north, as in the Bernese Oberland (Fig. 11).

Movement along these faults continued during the Late Jurassic, but, as the stratigraphy of the limestones of this age is yet imperfectly known, it is difficult to establish the direction of the movement. Things become clear again in

the Lower Cretaceous formations, most of which show different thicknesses on the two sides of the fault line and sedimentary breccias in the downset limb (Fig. 11). The elements of these breccias were embedded in a semiconsolidated state. One tectonic offset discovered by Schindler, earlier a normal fault with downthrow to the south, reversed its action in the late Valanginian and became a reverse fault, with a south-north thrusting effect (right-hand fault in Fig. 11). On the whole the fault movements during the Early Cretaceous resulted in a far smaller vertical throw than those of the Middle Jurassic or along the Liassic hinge lines, and their tensional nature is much less evident. This is, of course, expectable from the general tectonic history of the geosyncline.

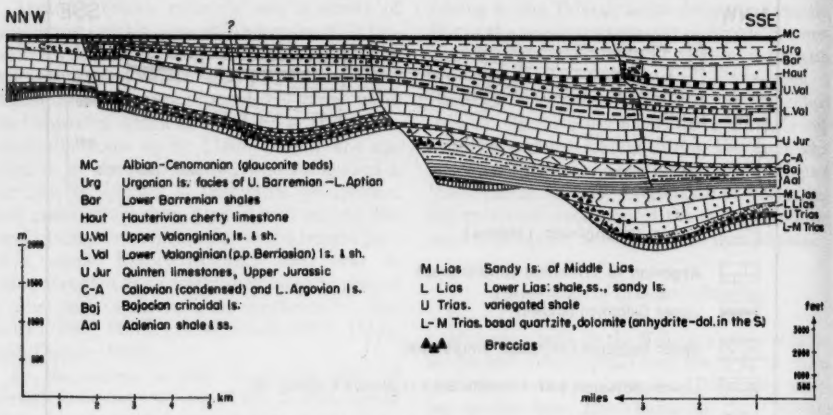


FIGURE 11.—PALINSPASTIC CROSS SECTION THROUGH THE TRIASSIC TO MIDDLE CRETACEOUS FORMATIONS IN THE GLÄRNISCH MOUNTAINS (HELVETIC NAPPES OF GLARUS ALPS)
After Schindler (1959), slightly modified. Vertical scale exaggerated × 2

Displacement along the Mesozoic faults of the Glarus nappes apparently took place by small, commonly repeated steps, and not simultaneously along the different faults. There is no evidence of the "orogenic phases" invoked by Günzler.

"Horst" characteristics of the Briançonnais realm.—The Briançonnais "geanticline" is the dominant paleogeographic feature inside the Penninic eugeosyncline of the western Alps, especially during Jurassic time (Figs. 1B, 2, 12). It was long considered a classical example of an asymmetrical embryonic fold (Gignoux, 1950; Moret, 1954). Staub (1951, p. 102) was probably the first to ascribe a horstlike structure to the Briançonnais rise, and the paper by Lemoine (1953) really started the revolution of its paleotectonic interpretation.

The Briançonnais realm encompasses not only the Briançonnais proper but also its northern prolongation in the Vanoise Mountains of Savoy, the internal part of the Préalpes Médianes, the Barrhorn Mountain group of Valais (Ellenberger, 1952b), and, very probably, the Schams nappes of central Graubünden. It separates the Subbriançonnais realm to the exterior (west or north) from the eugeosynclinal Piemont realm to the east or south. Its essential stratigraphic characteristics were outlined early in this paper. The sedimentary series is much thinner and less complete than in the adjoining belts, and up to the top of the Middle Jurassic (or, in the Prealps, to the Lower Cretaceous) it contains only shallow-water sediments.

Inside the Briançonnais facies belt there are generally no angular unconformities, despite great gaps. Over large areas, Bathonian or even Upper Jurassic limestones rest quite regularly upon Ladinian dolomites. Angular unconformities, at places very sharp, are limited to a very few localities. Two striking examples have been described by Tissot (1955, p. 159) from the vicinity of the Galibier Pass and by Debelmas and Lemoine (1957a) from the Peyre-Haute Mountains south of Briançon. Near the Galibier Pass the disturbance can be traced to a normal fault of Cretaceous age, and breccias occur in the downset eastern flank. On the whole, however, the sea bottom (or during part

of Early and Middle Jurassic time the land surface) on the Briançonnais rise was fairly level; the general configuration is that of a broad platform. In the type area of the Briançonnais it measured at least 50 km across the strike (Lemoine, 1953) and was affected by gentle undulations, which are expressed by the presence or absence of thin pelagic deposits of Oxfordian and Early Cretaceous age. Along the strike this platform extended from the Mediterranean to the Rhine.

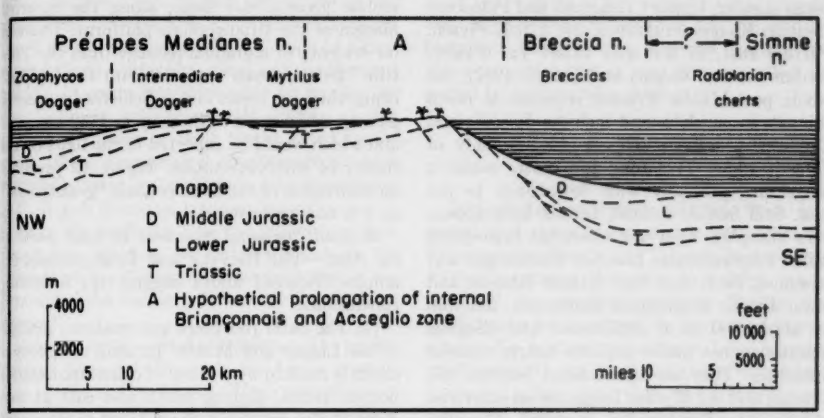


FIGURE 12.—VERY HYPOTHETICAL CROSS SECTION THROUGH SOME PREALPINE NAPPES, AT THE END OF THE MIDDLE JURASSIC
Vertical scale exaggerated × 2.5

The external (western or northern) margin of the Briançonnais belt is not very sharply defined. It is best exposed in the nappe of the Préalpes Médianes, where the Médianes rigides, to the southeast, represent the external part of the Briançonnais realm, and the Médianes plastiques are characterized by an almost complete stratigraphic sequence (Upper Triassic to Paleocene or lower Eocene). The limit between the two belts can be defined by the wedging out toward the southeast of the Upper Triassic,[11] the Middle Jurassic in *Zoophycos* Dogger facies, the nodular limestones of the Argovian, or the Lower Cretaceous cherty limestones (Figs. 2, 12; Pl. 2). These different facies limits do not coincide, and there is much crossing over. There are some small pre-Bathonian faults in this transitional zone, but no scarp breccias have been reported. At any rate, the external margin of the Briançonnais platform of Jurassic and Cretaceous time cannot be regarded as the front of a mobile embryonic fold or thrust fold.

The internal margin of the Briançonnais belt, its limit against the Piemont eugeosyncline, is much sharper and steeper. From north to south there are three regions where the passage between the two facies realms can be studied:

(1) In the Prealps, the Breccia nappe is probably derived from the most northwesterly part of the Piemont trough. We have seen that the Jurassic breccias of this unit were probably laid down east of a steep, tectonically active fault scarp. The components of the breccias are derived from the eastern marginal zone of the Briançonnais platform, mainly from its Triassic formations.

(2) The Vanoise Mountains of Savoy have recently become a key area of Alpine geology. Brilliant and painstaking work by F. Ellenberger (1949; 1958) has led to some amazing stratigraphic discoveries. In the western (external) part of the Vanoise, the sequence is very similar to that of the Briançonnais proper, and still more like that of the Médianes rigides. Discounting metamorphism, the sequence is quartzites of the Lower Triassic, thick lime-

[11] This definition is the most satisfying one (R. Trümpy, 1955b), as the presence or absence of the incompetent Upper Triassic beds has great tectonic consequences (basal thrust plane of the Médianes plastiques in the Carnian, whereas in the Medianes rigides, where there is little or no Upper Triassic, it is the Scythian).

stones and dolomites with gastropods and calcareous algae of the Middle Triassic, thin neritic formations of the Upper Triassic and lower Lias, Sinemurian to Bajocian missing, *Mytilus* Dogger, Upper Jurassic limestones, Lower Cretaceous missing, Upper Cretaceous and Paleocene Couches Rouges limestones, and a little Flysch. Farther east, in the zone called Val d'Isère-Ambin (*see* also Goguel and Lafitte, 1952), the whole post-Middle Triassic sequence is much thinner, incomplete, and includes Jurassic and Cretaceous breccias with boulders mainly of Triassic rocks. This zone apparently marks a hinge line, probably with downthrow to the east. Still farther inward, Liassic beds appear very abruptly. They are somewhat improperly called Prepiemontese Lias (*see* Ellenberger and Lemoine, 1955) and they include Rhaetic and lower Liassic fossiliferous limestones, followed by about 500 m of argillaceous and siliceous limestones, not unlike Schistes lustrés without ophiolites. They are intercalated between the Triassic and the *Mytilus* Dogger of an otherwise quite normal Briançonnais sequence. The relationship of the "Prepiemontese" Lias with the Piemontese Schistes lustrés is not yet established. The hypothesis that the Liassic formations of the eastern Vanoise represent a marginal facies of the Piemont Schistes lustrés is plausible. If it could be corroborated, this would prove the pre-radiolaritic (about Pliensbachian to Bajocian) age of the Schistes lustrés of the western Alps.

(3) East and southeast of Briançon, the typical Briançonnais zone and the Piemont Schistes lustrés come into direct contact; the latter constitutes an independent cover nappe. The discovery of the Acceglio zone (Debelmas and Lemoine, 1957b; Lemoine, 1957; Michard, 1959) is of capital importance for understanding the paleogeography of the western Alps. This zone forms the eastern prolongation of the Briançonnais zone, and it appears in windows below the Schistes lustrés nappe in the valleys on the Italian side of the chain. It is characterized by an extremely reduced stratigraphic sequence; a few meters of Upper Jurassic and Upper Cretaceous limestones rest upon the quartzites of the Lower Triassic or even on Permian formations. As the Jurassic and Cretaceous sediments are of pelagic facies, the Acceglio zone probably does not belong to a "cordillera," but rather to the upper (more westerly) part of the steep geosynclinal slope, whence most of the Triassic and Jurassic formations have crumbled off. Their debris may be found farther east, lower on the same slope, in the already mentioned microbreccias which appear in the upper Lias of the Gondran zone and in the Schistes lustrés on the western margin of the Piemont trough.

All these new results indicate a steep and mobile geosynclinal slope, along the internal margin of the Briançonnais platform. There is no evidence of a gradual passage from the "neritic" Briançonnais sequence into the bathyal comprehensive series of the Schistes lustrés—a passage which most authors up to 1949 assumed and which would be expected in the light of the theory of embryotectonics. Figure 12 presents an impression of the Briançonnais "geanticline" as it is now understood.

Mesozoic tensional structures in other parts of the Alps.—The Helvetic and Briançonnais examples discussed above suggest the following conclusions:

(1) The outer (northern and western) margin of the Liassic and Middle Jurassic miogeosynclines is marked by a series of contemporaneous normal faults, dipping south and east at medium angles. Fault blocks are tilted to the north and west.

(2) The Briançonnais facies belt appears, especially during Jurassic time, as a broad, gently undulating platform, limited by a steep, mobile scarp to the east and south.

The internal margin of the Piemont eugeosyncline can also be interpreted as a steep geosynclinal slope. Sediments of this slope should be sought in the cover of the lower Austroalpine nappes (Err and Bernina on Pl. 1). The works of Cornelius (1935), Roesli (1927), Staub (1948), and Stöcklin (1949) give a good idea of the Mesozoic stratigraphy of this area, in spite of controversy about the age of certain formations. The Mesozoic cover of the Dentblanche nappe, preserved at only one place in southern Valais, has the same general features (Hagen, 1948). The sequence is characterized by small thickness and abundance of sedimentary breccias of various ages. These formations were generally interpreted as consisting mainly of shallow-water sediments, laid down on and in front of the mobile geanticline at the front of the Austroalpine nappes. Several cordilleras and intervening troughs are distinguished by Staub (1917; 1948; 1958). The shallow-water origin of some formations in the lower Austroalpine nappes is supposed to be proved mainly by the abundance of breccias and by the incomplete sedimentary record, but it was concluded in the discussion of marine breccia formations that breccias of this type are not characteristic of any particular depth zone. The Upper Jurassic

uniformly shows the leptogeosynclinal pelagic radiolarian chert and aptychus limestone assemblage, and no neritic deposits younger than the middle Lias are known with certainty from the Upper Engadine. The realm of the lower Austroalpine nappes might in part correspond to the hinge zone between the Austroalpine platform (itself deeply submerged during much of Jurassic and Early Cretaceous time) and the Piemont trough. It seems not impossible to ascribe the very complex paleogeographic pattern of this area to a system of fault blocks tilted southward, like a symmetrical counterpart to those of the Helvetic realm. This is also the opinion of F. Roesli (personal communication).

As to the southern Alps, the influence of contemporary faulting on sedimentation in the Triassic of the Dolomite Mountains and in the Jurassic of the southern Ticino has already been stressed.

Remarks on the ophiolite problem.—The prevalence of tensional stresses in the Alpine geosyncline up to the Cretaceous affords a much easier explanation for the rise, intrusion, and outflow of ophiolitic magma than the theory of embryotectonics, which asked for compressional stresses during the whole Mesozoic. Fault and flexure zones along the tectonic scarps separating the platform realms (Briançonnais, including Subbriançonnais, and Austroalpine) from the Valais and Piemont troughs would provide the channels for the ophiolites, which spread out in the deeper eugeosynclinal belts but did not reach the platforms. The same phenomenon, on a much smaller scale, was earlier noted in the Middle Triassic of the Dolomites.

The mechanism of ophiolite intrusion is still somewhat controversial. Peridotite and gabbro bodies of laccolithic form are presumably intrusive. On the other hand, many spilites with pillow structures (and accompanied by tuffs) certainly represent submarine lava flows. But in the metamorphic central part of the Alps, where most ophiolites are now localized, it is difficult to arrive at a conclusion for individual ophiolite sheets.

Ophiolites are distributed mainly along two lines, following the internal margin of both Valais and Piemont eugeosynclines. Few or no volcanic rocks are found on the external margins of these troughs, in spite of the very pronounced faulting and warping along the scarp between the Briançonnais and Piemont belts. This systematic asymmetry—which was, of course, explained very elegantly by Argand's theory—may give a hint that ophiolite intrusion was early influenced by the beginning orogenic movements. Most ophiolites date from the (Early and Middle?) Cretaceous, and they were emplaced perhaps just when stresses in the geosyncline became reversed and crustal shortening first set in. By this time, the granitic layer below parts of the eugeosynclinal troughs may have become rather thin, but the gneissic basement cores of the Penninic nappes are there to prove that it was not altogether digested by subcrustal erosion.

Paleogeographic Pattern during the "Flysch Phase"

Early folding movements.—It has been suggested that the paleogeographic pattern of the Alpine geosyncline during Triassic, Jurassic, and Early Cretaceous time was to a large extent governed by tensional stresses. Basins and platforms were bordered by fault- and flexure scarps. No folding phases have been recorded with certainty up to the Early Cretaceous.[12] During this long time, shallow-water sediments were laid down on the platforms and in the bordering miogeosyncline, while Bündnerschiefer and "leptogeosynclinal" deep-sea deposits characterize the eugeosynclinal troughs, and the latter also a wide realm to the south. Breccias are localized along the geosynclinal slopes. It is certainly no coincidence that Flysch sedimentation, resulting from much more rapid erosional and depositional processes, starts as true orogenic folding movements begin.

Some of these premonitory Cretaceous fold movements may be listed here, without any attempt at completeness (*see* also Spengler, 1927):

(1) Pre-Cenomanian folding, especially in the western part of the higher Austroalpine nappes. M. Richter (*in* Kockel *et al.*, 1931, p. 97–104) has given an excellent description of this deformation, which produced rather gentle folds and normal faults with downthrow to the south.

(2) Turonian folding of the Devoluy Mountains in southern Dauphiné. This phase gave rise to well-defined east–west fold. (*See* Glangeaud and d'Albissin, 1958). Mercier (1958)

[12] The field evidence on which Paréjas' (1946) middle Liassic "Ferden phase" was founded is not confirmed by Baer (1959, p. 59). A "Neocimmerian" phase at the end of the Jurassic is mentioned in many published papers, but, although there was certainly some crustal unrest in various parts of the Alps at this time (Cornelius, 1937), proof of substantial folding has not been found.

gives the age of the disturbance as post-Cenomanian and pre-Late Turonian.

(3) Pre-Gosau (pre-Coniacian) orogeny of the Austroalpine nappes. The marine Gosau beds, containing a rich benthonic fauna, transgress thrust planes of the higher (older), décollement nappes of the northern calcareous Alps, especially in the Salzkammergut. The main thrusting is of Tertiary age, but the existence of Cretaceous strip sheets in this (originally very southerly) part of the Alps is proved. The Cenomanian is not known from the Salzkammergut, but farther west the unconformity between Cenomanian and Gosau is only slight; thus the exact dating of the pre-Gosau folding and thrusting is not established. In fact it is not even certain whether the pre-Cenomanian and the pre-Gosau disturbance are not identical; but tectonical analysis proves that there were two phases before deposition of the Gosau beds.

(4) Pre-Maestrichtian folding in the southern part of the Helvetic realm, indicated by the angular unconformity between the Wang beds (Fig. 10) and older formations down to the Upper Jurassic limestones. The folding dies out northward.

Important Cretaceous movements have certainly occurred in other parts of the Alps, but their exact age is less well established. In particular, the age of the deformation of the lowermost Penninic nappes (with Alpine granite gneisses) is not known, except that it lasted until the Miocene.

In Tertiary times, there were many orogenic phases, which may even overlap in time. Movements in the late Eocene ("Pyrenean phase") are especially important for Flysch sedimentation. The main phase of the western Alps falls about between early and middle Oligocene and coincides with the change from Flysch to Molasse sedimentation. The latest folding movements, which represent the main phase of the Jura Mountains, affected the uppermost Miocene or lowermost Pliocene formations. In the Pleistocene, there was considerable uplift and some faulting in the Alps.

Structural characteristics of Flysch cordilleras.—We may picture the Flysch cordilleras as chains of steep, disconnected islands, commonly rising immediately from rather deep troughs, but in places with narrow coastal or submerged platforms on which the larger Foraminifera and calcareous algae might prosper. Basement rocks cropped out in many places on these islands. As time and deformation went on, the more internal islands coalesced and formed a broad welt above sea level, while Flysch sedimentation continued, or started, in the more northerly and westerly parts of the original Alpine basin.

There is ample evidence of folding and even thrusting from the Cenomanian onward—that is, during the time when the different Flysch formations were laid down. We may certainly assume that the rise of the "Flysch cordilleras" was due to these tectonic movements, which resulted in crustal shortening. Debelmas (1957) has very lucidly demonstrated that these cordilleras are quite different from the older (especially Jurassic) "geanticlines," which seem to be horstlike structures caused by vertical and tensional movements, and that the compressional Flysch cordilleras did not necessarily develop on the site of older platforms.

Overlap of Flysch formations on folds and minor thrusts beneath is characteristic. The very nature of Flysch sandstones and breccias is evidence that the granitic and gneissic pre-Pennsylvanian basement lay bare and was eroded over large areas. The question whether Flysch transgresses over more important nappe thrusts, however, must be examined in more detail. For the lower Senonian Gosau beds of the eastern Alps, this fact is proved; but the Gosau beds, in spite of their comparative youth, do not show the characteristics of Flysch facies. In fact, they might rather be likened to a sort of older marine "Molasse".

The idea that the detrital matter of most Flysch formations, and especially the exotic blocks in the Wildflysch, was derived from advancing nappe fronts was first advanced by Schardt (1898). This would imply that the exotic boulders in the upper Eocene Wildflysch of the Ultrahelvetic nappes, for instance, might be of much more internal origin and might once have belonged to Penninic or Austroalpine nappes.

The gradual shifting of Flysch basins toward the outside of the Alps supports this hypothesis. Higher nappes ordinarily have older Flysch formations than do lower ones, although this is not an unbreakable rule. The upper limit of the Flysch in any belt, except the autochthonous one where it passes into Molasse, is fixed today by the base of a higher nappe. In some instances, cover nappes may have slid by gravity into a more external basin where Flysch was being laid down, and may actually have closed this basin; but, very commonly, the tops of Flysch formations were eroded before a higher unit overrode them. The folding progressed from

south to north (east to west in the French Alps), and the beginning and the end of Flysch sedimentation followed it; evolution of the Flysch. A source area north and west of many Flysch troughs excludes Schardt's hypothesis for these Flysch formations at any rate;

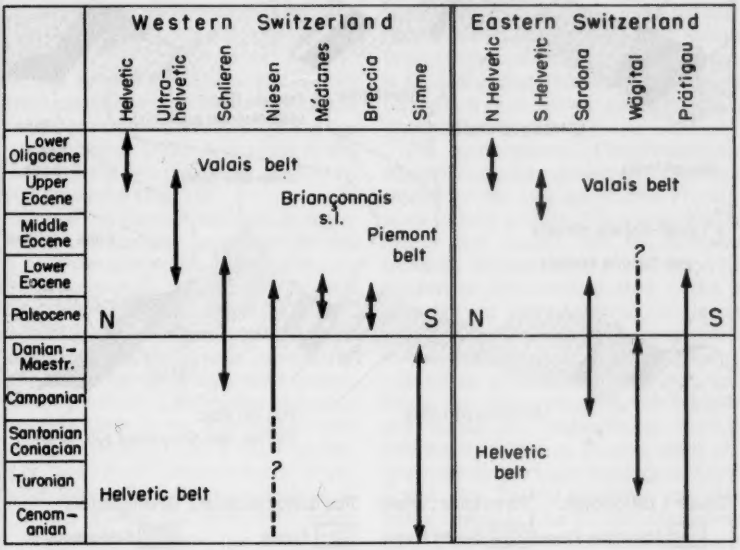


FIGURE 13.—TIME OF FLYSCH SEDIMENTATION IN DIFFERENT UNITS OF THE SWISS ALPS

Briançonnais realm in the broad sense lagged (Fig. 13).

Transgressive overlap of the Prätigau Flysch and its more southerly equivalents in Graubünden over most of the Penninic nappes, from the lower Penninic Adula up to the high Penninic Margna, was proposed by Leupold (1933) and was accepted by many geologists (Nänny, 1948; Staub, 1937; 1958; W. Ziegler, 1956). The present writer still feels uncertain as to the field evidence for this in the very complicated and moraine-covered key area around Tiefencastel, in central Graubünden.

Lugeon (1916) opposed Schardt's hypothesis on the origin of the exotic boulders in the Wildflysch and invoked a cordillera in the Ultrahelvetic realm itself. Most subsequent workers have followed this view (*e.g.*, Tercier, 1928; Leupold, 1942). It is confirmed by recent studies on current directions in several Flysch formations, which show for instance that the detrital matter in the Ultrahelvetic Gurnigel Flysch is derived from the northwest (Crowell, 1955; Hsu, 1960). A sand- and boulder-furnishing cordillera also lay north of the basin of the Prätigau Flysch and west of that of the Niesen nappes cannot have overridden the basin before the Flysch sedimentation, which commonly followed deposition of fine-grained pelagic sediments.

But it is wrong to state the problem of Flysch provenience as a choice between advancing nappes or cordilleras inside the geosyncline, as if the two were mutually exclusive. The material in the older and more internal Flysch formations is certainly derived to a large extent from cordilleras arising between the individual Flysch basins. The latest, north-Helvetic Flysch, however, received its detritus, including a large amount of volcanic rocks, just as surely from much more internal, south-Penninic nappes, which had already moved into proximity of the Flysch basin. In this respect also, conditions during the latest Eocene and early Oligocene were premonitory to those of the Molasse, where practically all the material is brought in from the rising Alps in the south.

In connection with the longitudinal filling of many Flysch troughs, Kuenen (1957b; 1958) has suggested a third possibility: an extra-Alpine source for the detrital matter, which would have been brought into the Flysch

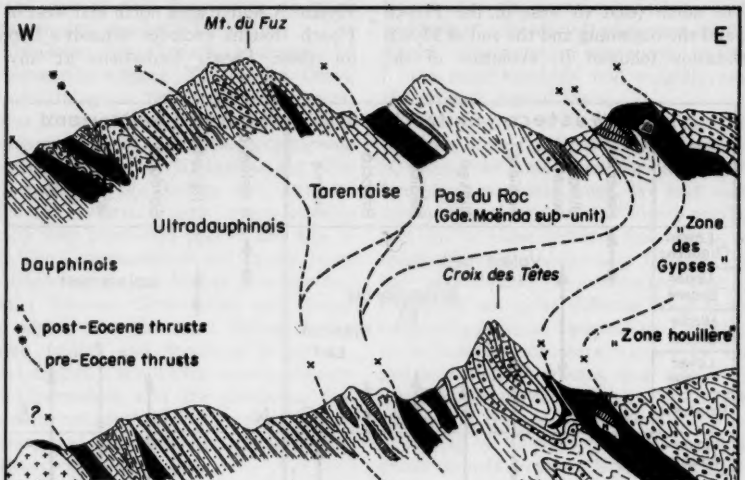


FIGURE 14.—CROSS SECTIONS THROUGH THE FRONTAL PART OF THE PENNINIC NAPPES IN SAVOY
After Barbier (1948). The lower section runs just north of the river Arc, the upper one about 11 km farther north.

troughs at one of their extremities. As Kuenen remarks, this possibility can probably not be applied to the Alps because of general paleogeography during times of Flysch sedimentation. Paleozoic and Precambrian siliceous rocks, which alone could have furnished the great amount of quartz and feldspar now contained in the Flysch, lay buried under carbonate rocks of Mesozoic age all over the country to the north and west of the Alps. Only certain areas in the Mediterranean, like the western part of Corsica, could be a possible external source—for instance, part of the Annot sandstones in southeastern France show a southerly source of detritus.

Alpine paleogeography of the final geosynclinal phase is commonly very difficult to unravel, especially because the exact origin of many flysch sequences that form cover nappes of their own is unknown or controversial. Flysch cordilleras also have a disconcerting tendency to vanish into thin air; thus, the peculiar Habkern granite which furnished great boulders to the Wildflysch and enormous masses of its characteristic greenish quartz and pink feldspar grains to the Gurnigel and Schlieren Flysch sandstones is not known in outcrop. The works of Kockel *et al.* (1931), Kraus (1932; 1942), Richter (1957), and Richter and Müller-Deile (1940) give a good idea of the complexity of troughs and cordilleras in the partly Ultrahelvetic but mainly Penninic Flysch zone along

the foot of the eastern calcareous Alps. These authors distinguish several cordilleras (especially the Rumunic and "Vindelician" ridge), in addition to the front of the advancing Austroalpine nappes, which all furnished detritus to the mainly Cretaceous Flysch and "pre-Flysch" basins.

Tarine and Arvinche cordilleras of Savoy.—Two Flysch cordilleras of different type in the western Alps have been illustrated by Barbier (1948) and Schoeller (1929). As a result of the Alpine orogeny the two cordilleras are now in immediate proximity (Fig. 14).

The internal or Tarine cordillera is represented in the Tarentaise nappe (nappe des Brèches de Tarentaise).[13] Its Lower Mesozoic sequence is incomplete: rather thin Triassic formations, followed by 50–200 m of Liassic limestones and locally some calcareous shales, which may belong to the Middle Jurassic. In the north, the Aroley limestones and breccias, with *Orbitolina* indicating Barremian or Aptian age, transgress these rocks. The Aroley beds are followed by probably still Cretaceous formations of "pre-Flysch" characteristics. South of the river Isère a thick Flysch succession of middle to late Eocene age rests with sharp angular unconformity upon the older formations down to the basement. The Tarentaise nappe thus has Cretaceous Flysch, or rather pre-flysch, in the north, and Eocene Flysch in the south (Barbier and Trümpy, 1955).

Thus a Tarine Flysch cordillera could not have existed in the Triassic. The thin and neritic Jurassic formations certainly mark a belt of shallower water, if compared with the thick, shaly sediments of the same age in the adjoining units of Savoy (Ultradauphinois slices to the west, Grande Moënda digitation of the Pas du Roc nappe to the east). In Valais, another digitation of the Tarentaise nappe, the Ferret zone, is present between the Ultrahelvetic (approximately Ultradauphinois) root zone and the Tarentaise proper. This unit has no Jurassic, but many hundreds of meters of Bündnerschiefer, probably of Early Cretaceous age, conformably underlying the Aroley formation. The Tarine "cordillera" of Mesozoic time was not stable; subsidence and uplift varied abruptly in time and in space. The unconformity at the base of the Cretaceous Aroley formation is much less pronounced than that below the Eocene Flysch, although, to judge by their components, the Aroley breccias overlay Triassic and basement rocks not far away. Only in the Eocene is there clear sedimentary proof of the existence of a cordillera; but there is no doubt that this cordillera arose over a belt which had shown great crustal unrest during Mesozoic times.

The more external Ultradauphinois zone, representing the Arvinche "cordillera," is capped by the Aiguilles d'Arves Flysch, which is up to 2000 m thick. The age of this Flysch is mainly late Eocene, but the base may be Lutetian. Barbier (1948, p. 120, 246; 1956) has beautifully demonstrated that it lies unconformably on its substratum of Jurassic to Paleozoic formations, and on the eastern border of the Pelvoux massif even on granite. This substratum possessed an imbricate structure before the transgression of the Flysch.[14] But this strong pre-Priabonian or pre-Lutetian deformation affects a Jurassic series of thick, monotonous shales and argillaceous limestones. Unlike the Tarine cordillera, the Arvinche tectonic ridge had no Jurassic antecedents. We know nothing about the Cretaceous history of the belt; formations of this age are missing and were probably eroded before deposition of the Aiguilles d'Arves Flysch.

This comparison is intended to show that Flysch cordilleras could develop on the sites of belts with quite different Mesozoic paleotectonic history.

Evolution of Paleogeographic Pattern

Ephemeral nature of facies belts.—Believers in embryotectonics long held that the troughs and ridges inside the geosyncline persisted practically throughout the pre-orogenic history of the Alps, and that a straightforward evolution led from the first faint bulges to the final great nappes. Modern stratigraphic research brings out a different story, and a bewildering instability of paleotectonic elements in space and time appears (Fig. 1). This instability is an essential feature of the Alpine geosyncline, more even than subsidence, and it is perhaps more pronounced here than in any other mountain chain.

[13] Barbier regards the Tarentaise nappe as an external unit of the Subbriançonnais belt, but its facies affinities are clearly with the Valais realm. The tectonic position of the unit of the Petit St. Bernard, with Schistes lustrés and ophiolites, is probably between the Tarentaise nappe and the Subbriançonnais proper. (*See* Elter and Elter, 1957.)

[14] The existence of rather violent tectonic movements preceding the deposition of the Aiguilles d'Arves Flysch certainly does not entitle it to be called "molasse" (Termier and Termier, 1957, p. 685).

The thick, monotonous, calcareous shaly groups of Schistes lustrés type, in particular, were often considered the result of continuous filling of rather deep, regularly subsiding, permanent troughs. These are the "séries compréhensives" of Termier, where sediments ranging without gaps from the Triassic or the Lias up to the Upper Cretaceous or to the Eocene were of uniform facies, in contrast to the varied and incomplete neritic Mesozoic sequence on the "geanticlines." The very valuable attempts at regional stratigraphic synthesis of the Bündnerschiefer by Jäckli (1941) and Nabholz (1945), for instance, are based on this working hypothesis.

But we have seen that the Schistes lustrés of the Cottic Alps probably represent only a comparatively short time span. The Piemont trough came into being in the Late Triassic, and, from the middle Lias onward, it subsided and received the Schistes lustrés group. The trough persisted right into the Cretaceous, but about in Bathonian time sedimentation became starved ("leptogeosynclinal"), probably by drowning of the surrounding platforms and by subsidence of the trough bottom to "oceanic" depths.[15] On the other hand, evidence shows that most of the Bündnerschiefer of the external, Valais eugeosyncline are of Cretaceous age, and that this trough was still incompletely developed during the Jurassic. In the Triassic, at any rate, the "trough" belonged to a rise that separated the German epicontinental Triassic basin, with its southern and very shallow marginal belt in the Helvetic realm, from the Briançonnais basin.

These results on the age of the Bündnerschiefer and Schistes lustrés are, to be sure, somewhat provisional. Still, it is fairly certain that environment did not stay constant in the two eugeosynclinal troughs during all of the Mesozoic era, or even during the whole of the Jurassic and Early Cretaceous. The thick sequences of the Bündnerschiefer-Schistes lustrés type were laid down fairly rapidly during rather short time, followed or preceded by episodes of arrested sedimentation. Many other, especially Paleozoic, geosynclines apparently received much thicker and more homogeneous shale-graywacke deposits.

If conditions were unstable in the eugeosynclinal troughs, they were even more so on the ridges, internal platforms, and scarps. The minor swells and furrows in mobile belts such as the Subbriançonnais-Médianes plastiques facies zone, paleogeographic elements a few kilometers or tens of kilometers broad, show repeated and rather irregular shifting of the axes of subsidence and of uplift (or "non-subsidence") during the Mesozoic. The monographs of Barbier (1948), Debelmas (1955), Jeannet (1912–1913; 1918), Peterhans (1926), and Schneegans (1938) describe striking examples. Others can be found in the Austroalpine and in the Helvetic-Dauphiné realm. (*See* especially Faure-Muret, 1955.) The mobility of the paleogeographic pattern is also expressed by the repeated penecontemporaneous erosion.

It is difficult to state any general rule about direction of displacement of the paleogeographic features. It is, of course, tempting to think of a gradual shifting of the axis of subsidence from south to north during the geosynclinal history (Cornelius, 1925; Frank, 1930). One might, for instance, list the belts with the greatest thickness for each system: Triassic in the southern Alps, Jurassic in the Schistes lustrés of the Piemont trough,[16] Cretaceous in the north-Penninic Bündnerschiefer, Eocene in the Ultrahelvetic Flysch, Oligocene in the Molasse of the foothills. But a more detailed analysis of Alpine stratigraphy reveals so many exceptions to this "rule" that skepticism arises as to its universal validity.

The author has recently summarized the essentials of Alpine pre-orogenic history (R. Trümpy, 1958; Fig. 1 of present paper), so it is not necessary here to discuss the vicissitudes of the paleogeographic pattern. It is essential, however, to examine whether the geosynclinal evolution is of one mold—as in Argand's theory—or whether we must distinguish different epochs, with basically different paleotectonic structure.

Relationship of late Hercynian structures to Mesozoic facies belts.—The region from which the Alps arose belonged to the central part of the great late Paleozoic fold belt of central and southern Europe. The history and structure of the Hercynian chains is extremely complex, and there is no reason to believe that any part of the future Alps was not affected by these movements. Circumstantial evidence seems to

[15] Lemoine's reservations (1959, and personal communication) as to this interpretation were mentioned under discussion of the "Schistes lustrés of the Cottic Alps."

[16] The Jurassic formations of the miogeosynclinal Dauphiné trough are in fact thicker, but subsidence in the Piemont eugeosyncline was stronger (arrest of sedimentation under "leptogeosynclinal" deep-sea conditions).

point to an Early Pennsylvanian ("Sudetic") date of the main deformation.

After this orogeny, and quite probably as a result of the isostatic disequilibrium created by it (Hsu, 1958), a number of elongated, fault-bordered basins developed in central Europe, four or five of them on the site of the future Alps. These basins were filled by very thick, detrital, mainly continental formations of Pennsylvanian and Permian age. Marine incursions of this age were essentially limited to the southeastern Alps. Volcanic activity was strong, especially during the early Permian, and furnished great masses of basic and acid lavas, tuffs, and ignimbrites. Fracturing and folding occurred at least twice thereafter, during the Pennsylvanian ("Asturic phase") and between the early and late Permian ("Saalic phase"), but these patterns were not parallel to the later Alpine structures. The divergence is clearly shown by a post-Eopermian and pre-Triassic syncline in the Aiguilles Rouges massif, which strikes N. 15° E., whereas the neighboring Chamonix "syncline," a narrow zone of much-aminated Mesozoic rocks separating the, Aiguilles Rouges from the Montblanc massif, runs almost exactly southwest-northeast. In the great basement thrust mass of the Austrian Alps, too, older structures commonly strike at right angles to the Alpine tectonic lines (Metz 1952).

By the beginning of the Triassic, the remnants of the Hercynian Mountains were levelled, and the clean, water-laid arkosic or quartz sandstones of the Scythian spread beyond the limits of the Paleozoic subsiding areas. In some of the late Hercynian basins, subsidence continued, resulting in much thicker Lower and Middle Triassic sequences than elsewhere. Most important of these is the "archaic Briançonnais geosyncline" (Ellenberger, 1949; 1950; 1958), which contains several thousands of meters of continental upper Paleozoic formations, up to 500 m of Lower Triassic quartzites, and up to 1000 m of Middle Triassic limestones and dolomites.

The question whether or not the Hercynian infrastructure had a major effect on the paleogeography of Mesozoic (or rather post-Mesotriassic) times is difficult to answer. Staub (1954; 1956) advocates a very marked influence of these structures, especially of fault lines and of the distribution of late Paleozoic granites. Strongly granitized belts would tend to behave as ridges during the later history. The stratigraphic evidence for this Hercynian inheritance is sometimes difficult to interpret,[17] as many of the nappes whose sedimentary history is well known are strip sheets of post-Triassic formations, and the nature of their pre-Triassic basement cannot be made out with precision.

The relationship of late Paleozoic to Mesozoic paleogeography has been especially well worked out by Ellenberger (1958) in the Vanoise Mountains. The axial part of the Lower Pennsylvanian to Middle Triassic "archaic Briançonnais geosyncline" became the non-subsiding Briançonnais platform of Mesozoic time. Upper Triassic and lower Liassic formations are only locally preserved on this platform, Pliensbachian to Bajocian ones not at all. (The "Prepiemontese Lias" of the eastern Vanoise may mark a temporary encroachment of the Piemont facies belt onto adjoining parts of the Briançonnais.) On the other hand, the Piemont belt to the east, which had received few sediments of late Paleozoic to Middle Triassic age, began to subside during the Late Triassic, notably in the Cottic and Ligurian Alps. In the Subbriançonnais belt to the west, the Upper Triassic formations are also thicker than in the Briançonnais, and they contain anhydrite beds of Carnian age. All through the western Alps, the Subbriançonnais units, including the Médianes plastiques, are stripped off along this incompetent anhydrite-bearing formation (Figs. 2, 13; Pl. 2). There are good reasons to believe (R. Trümpy, 1955b) that the complete section of the Subbriançonnais belt also had Lower and Middle Triassic formations of Briançonnais facies, although somewhat thinner than in the Briançonnais proper.

This is the phenomenon of "Reliefum kehrung" (Jenny, 1924), of inversion of subsidence. During the Triassic, the subsiding basin became consolidated and developed into a nonsubsiding platform ("geanticline"), and new geosynclinal basins were born outside the ancient ones. The Piemont eugeosyncline started to subside in the Carnian, at first as a shallow depression; but it deepened during the Lias to reach oceanic depth during the Middle Jurassic (assuming a pre-radiolaritic age for the Schistes lustrés). The Ladinian-Carnian transition was certainly a crucial moment in the early history of the Alps. It marks the end of the last remnants of Hercynian structures

[17] As far as the Glarus Alps are concerned, the writer feels unable to adopt the views of his teacher Staub (1954).

and the birth of the Alpine paleogeographical units.

There might be a Hercynian heritage in post-Triassic paleogeography, but the correspondence of the two structures is inverted, the negative elements of one epoch becoming the positive ones of the next. The present writer has reservations about this as a general rule, however. The spatial coincidence of the "archaic Briançonnais geosyncline" and the Mesozoic (essentially Upper Triassic to Middle Cretaceous) Briançonnais rise in the Vanoise Mountains may be accidental. The two structures are not parallel; farther north, in the Valais-Prealps cross section, the post-Carnian formations covering the main part of the pre-Carnian subsiding basin are of Subbriançonnais, not of Briançonnais, facies, and in Graubünden the Upper Jurassic limestones of the Sulzfluh nappe, which almost certainly represents the prolongation of the Briançonnais facies belt, rest directly upon granite. This angular divergence of Hercynian structures and Mesozoic facies lines excludes, in our opinion, a far-reaching genetic influence of the former on the latter.

Relationship of Flysch cordilleras to Mesozoic platforms.—Emphasis was earlier placed on the essential paleotectonic difference between the Jurassic and Early Cretaceous topography on the one hand and that of "Flysch times" on the other. It remains to inquire whether the two types of structures are interrelated.

Again, the fate of the Briançonnais platform is highly significant (Figs. 1, 2). It came into being in the Carnian and had its heyday between the Pliensbachian and the Bathonian (Fig. 12), when most of it was dry land, while great masses of shales and of argillaceous and silty limestones were laid down on either side (Lias and lower *Zoophycos* Dogger of the Subbriançonnais-Médianes plastiques, Schistes lustrés of the Piemont trough). It is still well expressed during the later Middle Jurassic, by the shallow-water facies of the *Mytilus* Dogger. In the Late Jurassic, the difference of facies in the Subbriançonnais belt becomes slighter, whereas the deeper, leptogeosynclinal Piemont belt was still quite distinct. Apparently, the Briançonnais platform was again better marked during the Early Cretaceous. Sediments of this age are missing on it, whereas there are pelagic limestones with chert layers on either side. Lemoine (1953) has shown, however, that, in the type area of the Briançonnais at least, pelagic conditions prevailed over the whole belt, and sedimentation was inhibited only by currents and paleogeography. Finally, the pelagic, although not necessarily deep-water, limestones of the Cenomanian to Paleocene Couches Rouges formation are practically identical in the Briançonnais realm, in the Subbriançonnais belt, even in the Falknis nappe of Graubünden, and also in the parts of the Piemont realm which lie next to the Briançonnais (Breccia nappe in particular).

While the thin Couches Rouges were being laid down over a Briançonnais platform which had more than doubled its breadth and whose boundaries with the adjoining facies belts had become quite vague, Flysch was already being formed in more distant belts (Fig. 1). These are (1) the Valais realm to the north (Cretaceous formations of Schlieren Flysch, Niesen Flysch, Prätigau Flysch, etc.), (2) the internal parts of the Piemont and probably also the external parts of the Austroalpine realm to the south and east (Simme Flysch, Helminthoid Flysch of the southwestern Alps, etc.). Detrital matter was trapped by these troughs on either side of the Couches Rouges ("Briançonnais") platform, and only in the late Paleocene and early Eocene was a little Flysch spread over it.[18] Except at one place (Wegmüller, 1953, p. 44), no instance is known where Flysch transgresses pre-Cenomanian formations of the Briançonnais realm proper. This platform, by far the most prominent feature inside the Penninic geosyncline of the western Alps during Jurassic time, became obsolete by degrees, and no longer played an important role during the early phases of the Alpine orogeny. This fact is surely quite incompatible with the concept of embryotectonic prefiguration of nappes.

This picture is a very general one. In the Helvetic realm, too, the faults and flexures of "Mesozoic," tensional type were most active during the Lias and the earlier Middle Jurassic. During the Late Jurassic and the Early Cretaceous, displacements along these faults were much smaller. Flysch sedimentation again develops quite independently from the ancient structures; basins and cordilleras wander gradually outward.

This does not mean that Flysch cordilleras nowhere developed on the sites of former ridges,

[18] Both the "Plattenflysch" of the Prealps (Delany, 1948; Klaus, 1953; Twerenbold, 1955) and the Helminthoid Flysch of the southwestern Alps (Lanteaume, 1957), which were long thought to belong to the same tectonic unit as their substratum of Briançonnais facies type (including Paleocene Couches Rouges), are of Cretaceous age and of more internal origin.

platforms, or scarps. The previously discussed Eocene Tarentaise cordillera certainly formed on the site of a belt that had been rather strongly deformed during the Mesozoic. The eastern margin of the Briançonnais belt, which was active during both the Early Jurassic and the Late Cretaceous, may be another example (Lemoine, 1957). The point to be stressed is that Flysch cordilleras—an early expression of the compressional tectonic forces by which the Alpine nappes were built—could arise wherever the causal stresses obtained, and that their location was little influenced by the disposition of the Mesozoic troughs and rises, many of which had become indistinct before Flysch sedimentation set in.

Obliquity of Paleogeographic Units

General statement.—The hypothesis of embryotectonics implied that there was a direct, genetic link between paleogeographic belts and tectonic units. The great nappes and, concomitantly, the major facies zones were considered to have an almost unlimited extension along the strike of the chain. This geometrical and paleogeographical constancy furnished a double handhold for identifying and comparing individual nappes, often in great detail, from one end of the chain to the other. The works of the great masters of Alpine tectonics, such as Argand, Lugeon, Kober, Staub, and Termier, are full of discussions on the correlation of nappes observed in the different transverse sections between Vienna and Genoa, or even beyond to the Carpathians and to Corsica.

This "cylindrist" dogma was vigorously attacked by other geologists. To these "festoonists," the most remarkable of whom was Gignoux, the Alpine paleogeographic units form festoons of limited extent, relieving each other along the strike of the chain and of the geosyncline. To adherents of this line of thought, the original basin of the Niesen Flysch, for instance, was never much longer than the 60 km along which the Niesen nappe is now preserved (Lugeon and Gagnebin, 1941, p. 51). Criticism of the Argandian edifice was certainly necessary, and it is only natural that it overshot its mark. Considering the Alps as a regular geometric body and as a chaos are two comfortable, but not necessarily very productive, working hypotheses.[19]

The question of obliquity and limited extent of paleogeographic units must be examined from two directions: nonparallelism of facies zones of different ages, and nonparallelism of paleogeographical and structural entities.

Crossing over of facies belts.—The instability of Alpine paleogeographic features implies that facies belts of different ages can run at angles to each other. This is notably the case for the late Hercynian basins and the Mesozoic geosynclines. But even during the Alpine preparatory phase, crossing of the isopic lines (lines of equal facies, Arnold Heim, 1908) for different formations can be easily demonstrated.

Tectonic structures in the Helvetic nappes of Switzerland trend in general west-southwest to east-northeast. As already mentioned, Liassic formations are absent from the originally northern part of the Helvetic realm, and they set in very rapidly along a line which is interpreted as a contemporaneous flexure. This line runs about east-west, with some embayments. It is now seen in the higher Helvetic nappe of eastern Switzerland, in the cover of the southern flank of the Aar massif in central Switzerland, and in the Chamonix "syncline" between Aiguilles Rouges and the Montblanc massif in western Switzerland. The axis of the Liassic Dauphiné geosyncline, with its thick formations of dark shales and shaly limestones, runs about parallel to the marginal scarp and can be followed from the Ultrahelvetic cover of the Gotthard massif to the Helvetic nappes of the Bernese Oberland, to the Subhelvetic Morcles nappe in the Rhone Valley, and finally to the Subalpine chains of Savoy. Isopic lines for the Middle Jurassic formations have a similar orientation, whereas those for the Cretaceous run more nearly parallel to the structural trend. The Eocene facies belts are arranged quite otherwise and run about southwest-northeast when plotted on a palinspastic map (Boussac, 1912; Leupold, 1942). They form angles up to 35° with the tectonic zones and up to 45° with the Liassic facies belts. The northwestern limit of the middle Eocene transgression, for instance, touches the Rawil pass, between Valais and the Bernese Oberland (high Helvetic Wildhorn nappe), the Pilatus near Lucerne (border chain, belonging either to the front of the Wildhorn-Drusberg or to the back of the lower Axen nappe), and the Kistenpass between Glarus and

[19] The "festoonists" are often also called "Little-Alpians" (Swift, Voyage to Lilliput), meaning that they have a tendency to minimize the amplitude of nappes and the amount of crustal shortening. The author adheres to the sect of noncylindrist Big-Alpians, which probably includes the majority of French and Swiss Alpine geologists of the younger generation among its members.

Graubünden, in the autochthonous cover of the eastern Aar massif.

The limit between the Briançonnais and Subbriançonnais belts in the Median Prealps nappe of western Switzerland furnishes another example of crossing over of facies belts. The Briançonnais-Médianes rigides realm is readily distinguished from its northwestern neighbor by its lack of Upper Triassic, lack of Lias, absence of Middle Jurassic or representation only by Mytilus beds, lack of Argovian nodular limestones, and lack of Lower Cretaceous (Fig. 2). The lines marking these different facies limits cross each other in a complicated and unsystematic manner; as a result it is impossible to assign the klippen of central Switzerland (tectonic outliers of the Médianes nappe) to a definite facies belt of the same nappe in western Switzerland (Peterhans, 1926; Weiss, 1949). The Mithen, east of Lake Lucerne, for instance, show a very "northerly" facies of the Upper Triassic and the Middle Jurassic, but also "Briançonnais" characteristics, such as absence of Lower Jurassic and Lower Cretaceous formations.

The review by de Sitter and de Sitter-Koomans (1949) on the Bergamask Alps also contains very suggestive maps showing the obliquity of facies zones in the Permian and Triassic of the southern Alps.

Nonparallelism of paleogeographical and structural units.—The cylindrist working hypothesis that the major Alpine nappes did not essentially change their geometrical characteristics along the strike of the chain has been challenged on tectonic grounds. The axes of microfolding show a very complicated pattern in areas like the Vanoise (Ellenberger, 1958) or the gneiss nappes of the Ticino (Wenk, 1956). The major structural units, although they do not in all cases conform to the trend of the microfolding, also show strong deviations from normal strike. This makes the detailed parallelism of nappes on a purely geometrical base very problematical, but it seems possible to give an approximate correlation of units on either side of the transverse Ticino uplift. Paleogeographical evidence, of course, has been widely used for nappe correlation, but the two methods should be kept apart, as we cannot *a priori* assume the parallelism of structural and paleogeographical units.

In most of the French Alps, the Subbriançonnais realm appears to be directly contiguous to the Dauphinois-Ultradauphinois miogeosyncline, and most workers agree that there were no other zones between the two (*e.g.*, Gignoux, 1950, p. 311). North of the river Arc, in Savoy, a new element not represented in the classical section of the Durance Valley appears between the Ultradauphinois and Subbriançonnais zones. This, the Tarentaise nappe, is the most southwesterly evidence of a facies belt for which the author (1955b) reintroduced Haug's term of "Valais geosyncline" or, better, Valais realm. This belt broadens to the northeast and takes in more elements: the Ferret zone to the outside and the Schistes lustrés of the Petit St. Bernard to the inside of the Tarentaise nappe. The most external digitation of the Subbriançonnais units (Grande Moënda), with its shaly facies of the Lower and Middle Jurassic, may mark the transition between the Subbriançonnais and the Valais facies belt. The original breadth of the latter in Valais may have been of the order of 100 km.

Farther east, in Graubünden, the Valais facies belt is much broader, more complex, and distinctly eugeosynclinal. All the Bündnerschiefer and Flysch masses of northern and central Graubünden belong to it. The Subbriançonnais belt loses its individuality between Ticino and Lake Lucerne. Equivalents of the Briançonnais facies belt are found only in the Sulzfluh nappe overlying these schist groups and in the Schams nappes (especially the Kalkberg and Marmorzone units) south of them. These complicated décollement nappes lie near the front of the Suretta basement nappe, which is believed to correspond to part of the Grand St. Bernard nappe of the western Alps. If they belonged to the original cover of the Suretta nappe there would still be concordance between basement nappes and paleogeographical zones, in spite of the great broadening of the north Penninic Valais realm. But those who have worked on the Schams nappes (especially Staub, 1937; 1958; Streiff, 1939; Neher, personal communication) deny this possibility and generally agree on the derivation of most Schams units from the high-Penninic Margna nappe, which nobody would think of correlating with the Grand St. Bernard nappe. In this case, the discrepancy of facies belts and major nappes (basement nappes) is total, and the comparison of tectonic units by paleogeographical evidence becomes impossible. Between Savoy and Graubünden, the Briançonnais platform would run across almost the whole breadth of the Penninic geosyncline. In the French–Italian Alps, the eugeosynclinal Piemont trough lies inside (east) of the Briançonnais rise. In Valais and Ticino, there are two eugeosynclines, of about equal importance,

on either side of it. In Graubünden the external Valais trough is by far the more prominent; if the Schams nappes are derived from the Margna, equivalents of the Piemont belt are present only in the Platta (= Arosa) cover nappe. If the basement nappes east and west of the Ticino transverse uplift can be correlated at all, which many authors doubt, it is only on purely geometrical grounds.

We should probably distinguish more rigorously between the thick nappes with basement cores and the thin décollement nappes (strip sheets) consisting only of Mesozoic and lower Tertiary formations. There has long been a tendency to assign each décollement nappe to a particular basement nappe, and to regard the former as a kind of frontal prolongation of the latter. The two phenomena are probably of different order, however, and came into existence at different times. It becomes apparent that in many instances the Mesozoic cover was stripped off its basement before the latter became involved in the major nappe-building processes, at a rather early state of the Alpine orogeny.

This is evident in particular for the Median Prealps, which originated in an area that was later affected by Alpine metamorphism. They must have left this area and moved northwestward, bearing the Simme nappe on their back, before metamorphism set in, and before the "lid" of the Austroalpine Dentblanche nappe, with its mighty basement core, covered the Penninic zone of the western Alps. Metamorphism is contemporaneous with or slightly later than the main folding and thrusting of the Penninic gneiss nappes.

Structural evidence shows that the Breccia nappe and especially the Niesen nappe were the latest units to take their place in the Prealpine edifice, the Niesen nappe even after the formation of the Helvetic nappes (Badoux, 1945). These two nappes are the only ones that have any (although small) remnants of pre-Triassic formations, and the basal beds of the Niesen nappe are the only formations of the Prealps that bear the stamp of Alpine metamorphism.

For the Austroalpine cover nappes of the northern Calcareous Alps, the case is somewhat different. Initial stripping off of the cover began during the Cretaceous pre-Gosau phase, but the main movement took the décollement nappes, together with their basement of the Silvretta-Oetztal nappe, over the Penninic area. Additional independent gliding of cover nappes occurred during this Tertiary deformation (Fallot, 1954).

The basal shear planes of décollement nappes are governed largely by the facies of the Mesozoic formations, especially by the incompetent anhydrite members of the Triassic. Thrusts will follow incompetent formations for a long distance and then cut at a larger angle through competent ones. Facies changes affecting the competency of the rocks—anhydrite to dolomite, or shale to limestone—will strongly influence the structure of cover nappes (for instance, Fallot, 1953; Helbling, 1938; Lemoine, 1952; de Sitter, 1939; R. Trümpy, 1955a). Because the localization of thrust planes is determined by facies, cover nappes are in some instances more nearly parallel to the paleogeographic zones than are the basement nappes.

The two phenomena—the stripping of the cover and the formation of "gneiss nappes"—are not necessarily directly related. Many basement units lost their cover at an early stage, either by shearing off or by erosion during the "Flysch phase" of geosynclinal history. In some it has been replaced by a "false cover" that moved forward from more internal zones ("remplacement de couverture" of Ellenberger, 1958). As far as the large nappes with basement cores are concerned, there is no simple genetic link between Mesozoic paleogeography and present-day nappe structure. As the author has already observed (R. Trümpy, 1958), "this does not make things easier for the Alpine geologist, and still less so for his extra-Alpine reader."

Conclusions

Five Epochs of Alpine History

The structural history of the Alps can be divided into five epochs, each of which has engendered its own set of sedimentary rocks.

First: The Hercynian post-orogenic and anorogenic phase (would be included in the "pre-orogenic" phase of most authors) started with the subsidence of elongated, commonly fracture-bounded basins on the Hercynian chain after its main deformation. During the Pennsylvanian and Permian, these basins were filled by detritus from the Hercynian Mountains. When they were levelled, by the beginning of the Triassic, sedimentation spread beyond the "epieugeosynclinal" basins, but they remained the seat of stronger subsidence than the other areas, especially in the case of the "archaic Briançonnais geosyncline." Another belt with rapid subsidence lay in the

southeast (South Tyrol, highest Austroalpine nappes); possibly its evolution was advanced beyond that of the main part of the future Alps.

Second: Alpine pre-orogenic epoch ("Schistes lustrés times"). In the western Alps, the change from late Hercynian to early Alpine paleogeography can be dated precisely (Ladinian-Carnian transition), but the Triassic as a whole was a transitional period. The Piemont geosyncline commenced in the Late Triassic as a gently downbuckled basin, but from the middle Lias onward it became a deep and rapidly subsiding furrow. The paleogeography of Early and Middle Jurassic time shows a marked contrast between troughs and nonsubsiding but instable platforms. The Piemont Schists lustrés accumulated in the troughs. Breccias are prominent along geosynclinal scarps. The external miogeosyncline received even thicker sediments than did the eugeosynclinal belts. In the Late Jurassic, very little detrital matter was brought into the geosyncline; the Piemont trough entered into its "stage of vacuity" (Aubouin, 1959) and received only thin deep-sea deposits. Most of the Bündnerschiefer of the Valais trough are of Cretaceous age. The evolution of the Valais and Piemont eugeosynclinal belts was obviously not concordant.

The structural pattern of this pre-orogenic epoch is governed by vertical movements, perhaps even tensional stresses resulting from a stretching of the crust. Normal faults and flexures mark the borders between platforms and troughs. During the Cretaceous, ophiolites found their way along these fracture belts, perhaps at the very time when compressional stresses set in.

There was probably a period of less contrasting submarine topography between the phase of strong vertical movements (Early and Middle Jurassic) and the onset of the tangential movements (Middle Cretaceous to Eocene).

Third: Early synorogenic or cata-orogenic phase ("Flysch times"). From the Cenomanian or, locally, the Early Cretaceous onward new ridges ("cordilleras") arose in the Alpine geosyncline. Their location shows no simple relationship to the platforms and troughs of the preceding epoch. These cordilleras rose above sea level and furnished great masses of detritus to the remaining troughs on either side, which were rapidly filled by Flysch sediments. Flysch sedimentation started in the Austroalpine realm, in the Piemont and Valais troughs, but moved outward (northward and westward) in course of time and finally encroached upon the Briançonnais platform, the Helvetic miogeosyncline, and even the nongeosynclinal belt beyond it (Fig. 13). In these areas, downbuckling preceded the arrival of the flood of detrital matter; pelagic formations invariably underlie Flysch.

Flysch cordilleras show evidence of folding and thrusting. While compressional stresses grew stronger, the outflow of ophiolites ceased. All through "Flysch times," orogenic deformation went on. The earliest nappes of the eastern Alps are Late Cretaceous; those of the western Alps are late Eocene. By the time the youngest (early Oligocene) Flysch was laid down over the Helvetic realm, most of the Alps already was above sea level. In any particular Flysch basin, the major thrusting movements put an end to Flysch sedimentation, but in some places Flysch is transgressive on folds and minor thrusts.

Fourth: Late synorogenic or epi-orogenic phase ("Molasse times"). The last remnant of the geosyncline shifted to the external border of the chain; the narrow remaining arm of the Alpine sea became shallow and brackish and was finally filled by detritus from the rising Alps. Strong subsidence of the Molasse "exogeosyncline" goes along with the upheaval of the Alps. The major thrusts are formed, metamorphism and granitization (on a smaller scale than in many similar chains) take place. Erosion attacks the chain down to several kilometers below the original structural surface. From late Oligocene to middle Miocene, the Alps formed a chain of high mountains, but they were not more than a hilly tract of country by the beginning of the Pliocene. The present morphology, especially of the western Alps, is the product of Pleistocene uplift and erosion. Post-orogenic volcanic activity started in the Eocene south of the Alps, in the Miocene to the north.

Fifth: Compressional deformation has almost ceased (Pliocene and Pleistocene), but vertical movements continue, perhaps with some displacement along strike-slip faults.

Remarks on Geosynclinal Nomenclature

The author had intended to devote a section of this paper to geosynclinal nomenclature, but while it was being written Aubouin (1959) published an excellent and well-documented review of the problem. Its treatment also by Dunbar and Rodgers (1957, p. 312) and by King (1959, p. 54) shows that a consensus is imminent. The proposed restriction of the term "geosyncline" to the orthogeosynclines of

Stille would certainly clarify things (*see* also Glaessner and Teichert, 1947) and would be approved by most European geologists, who for more than half a century have used the term only in connection with the mio- and eugeosynclines from which folded mountain belts arose.

If, for historical reasons, it is necessary to keep Dana's broad definition—any elongated basin which received an abnormally great thickness of sedimentary formations, as compared to its surroundings—then it becomes necessary to distinguish a large number of different kinds of "geosynclines" of quite different geotectonic significance, by prefixes or by some other means. Most geologists probably do not favor using the word "geosyncline" for shallow basins of almost circular outline ("autogeosynclines") or for fault-bordered troughs that lie outside the fold belts and contain primarily continental deposits ("taphrogeosynclines"). Nevertheless, one of Dana's original examples of a geosyncline was the Triassic "taphrogeosyncline" of Connecticut.

Classifications of course nearly always provide too rigid a set of boxes for the great number of natural variations observed. Kay (1947; 1951), who has been criticized for setting up a large number of categories for geosynclines, has done much to show how the different types grade into each other.

Even if one restricts the term to orthogeosynclines, it is commonly difficult to decide what is a geosyncline and what is not. The Molasse foredeep ("exogeosyncline")—not a geosyncline by the author's standards—gradually develops out of the Alpine geosyncline, which shifts and is filled. Distinction between eugeosynclines and miogeosynclines also may be controversial. Volcanism cannot be used as the only criterion: the Triassic of the Dolomite Mountains is a good example of miogeosynclinal deposition with volcanic material, the Bündnerschiefer of the Prätigau exemplifies eugeosynclinal sedimentation without volcanism. Should Flysch formations be classed as miogeosynclinal or eugeosynclinal? They develop over mio- and eugeosynclines alike; they have no volcanic materials, but conditions of relief and of contemporaneous mobility approach those of true eugeosynclines.

Another point concerns the thickness of accumulated sediments, considered as the essential characteristic of geosynclines by Hall and Dana. Many otherwise quite typical geosynclines received little sediment over long intervals. If this is generally recognized, the term "leptogeosyncline" becomes useless, and the author will gladly lend a hand to bury his own brain child. The term is necessary only as long as the historical definition of geosynclines is upheld.

Many geologists—certainly including the present author—use the term "geosyncline" in a sloppy way for features of quite different order of size and complexity (Thetys, Alpine, Penninic, and Piemont geosynclines). The individual troughs which make up a geosyncline should be distinguished under other names, such as secondary geosynclines (Haug) or intrageosynclines (Tetiaev; *see* Aubouin, 1959, p. 156). Trying to be unmistakably precise, however, can make geological language heavier and heavier; doubt naturally arises whether it is a material advance to science to call the Piemont belt in the Upper Jurassic a meso (?)-ortho-eulepto-intrageosyncline.

Peculiar Characteristics of the Alpine Geosyncline

The Alpine geosyncline can certainly not serve as a suitable standard specimen of "The Geosyncline"—leaving aside the historical "type" specimen in the western Appalachians. First of all, reconstruction of Alpine paleogeography is much too difficult, and over large tracts very doubtful, because of the great structural complications, metamorphism, and primary lack of fossils in many formations. No really secure synthesis can be given until we know more about the stratigraphy of the Schistes lustrés and before settling the controversial points regarding the origin of several important cover nappes—to mention only Helminthoid Flysch, Simme, Niesen, Schams, and the Flysch nappes along the border of the eastern Alps. On the other hand, the Alpine geosyncline differs materially in several ways from many others.

Compared with those of many other (especially Paleozoic) mountain chains, sedimentary rocks of the Alpine geosyncline are very calcareous. This is normal for the miogeosynclinal belts, but even the eugeosynclinal Schistes lustrés have a high proportion of calcium carbonate. Typical graywackes are absent in the Schistes lustrés and not very abundant in the Flysch. Pre-orogenic ophiolitic volcanism is relatively modest.

Sediment accumulation in the Alpine basins was less pronounced than in many other geosynclines (rates cited in Kay, 1955). During long intervals, the eugeosynclinal troughs and

areas to their interior received only very thin deposits. This "period of vacuity," still more evident in the Tethyan geosynclines around the eastern Mediterranean, in Greece and Turkey, may be due to the existence of paraoceanic conditions during the second half of the Jurassic and the beginning of the Cretaceous. Even the Flysch sequence of the Alps looks modest, compared with that of the Carpathians or of the northern Apennines, but this is perhaps because the Alps are more deeply eroded; in the late Oligocene they may have resembled the Carpathian and Apennine "Flysch mountains," to judge from the pebbles in the Alpine Molasse conglomerates. Also, where the deepest parts of the northern Apennines are exposed, in the Apuanic Marble Mountains, rocks and structure look decidedly "Alpine."

The essential feature of the Alpine geosyncline is perhaps its instability. Instead of a permanent trough, in which great masses of sediments and volcanic deposits were heaped up, there is a bewildering succession of troughs and rises, some of them quite small and quite ephemeral. Crustal unrest, rather than subsidence, is the hallmark of the sedimentary history of the Alps.

Does this very turbulent pre-orogenic history reflect the same deep-seated causes as the extraordinary tectonic complications of the Alps? It seems that the nappe structure of the Alps is not an isolated, but certainly an extreme, case. This complicated structure severely limits reconstruction of Alpine paleogeography. We have learned a lot of new facts in the last 10 years, chiefly concerning the western Alps; but much remains to be done before a coherent paleogeographical synthesis can be attained. For this reason, study of simpler mountain chains will perhaps contribute more significantly to the genesis and the life history of geosynclines than will that of the Alps. Nevertheless, the Alps are certainly significant in that they afford some insight into the very complex past of a very complex chain. Some points now fairly well established are:

(1) An interval of only slightly contrasting submarine topography precedes the birth of the Alpine geosyncline. Subsidence of the late Paleozoic "epieugeosynclines" can be ascribed to isostatic compensation following Hercynian folding, but not so the formation of the Alpine geosyncline itself, which followed the last preceding folding by about 30 million years.

(2) During the earlier stages of Alpine geosynclinal history slightly subsiding platforms and strongly subsiding troughs were limited by fault scarps. Movements along these scarps are vertical and denote tensional stresses; they were strongest during the Lias and the earlier Middle Jurassic.

(3) Deep-sea sediments, especially of Middle Jurassic to Early Cretaceous age, are widespread in the eugeosynclinal belts and also farther south, in the Austroalpine and south Alpine realms. At that time the Alps may have represented an incipient or threshold ("monoliminary") orogenic belt (Glangeaud, 1957).

(4) The cordilleras of the Flysch stage were steep island arcs, engendered by compressional stresses and not directly related to the earlier platforms.

(5) On the whole, paleogeographical units are rather ephemeral and subject to repeated shifting.

(6) There is no simple relationship between rises inside the Mesozoic geosyncline and Tertiary (or Late Cretaceous) nappe structures.

This last point needs one final comment. The author has insisted that there is no simple, direct, readily understood relationship between paleogeographical elements and major nappes, such as was provided by the theory of embryotectonics. Some relationship evidently does exist, but it is not simple and has yet to be worked out. The different structural patterns which succeeded one another in time and space during the preparatory and paroxysmal phases of Alpine history are certainly, in some complex way, all expressions of one great process. The nature of this process is, in the main, a geochemical and geophysical problem, to the solution of which the field geologist and stratigrapher can contribute mainly by presenting and co-ordinating the facts, and by demanding that a theory of mountain building accord with and account for them. This the author has attempted to do in greatly condensed form, and from the works of several generations of Alpine geologists.

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5. GEOLOGY OF SAINT JOHN, NEW BRUNSWICK. A. O. Hayes and B. F. Howell. 1937. 146 p., 9 pls., 4 figs. $1.50.
7. ATLANTIC SUBMARINE VALLEYS OF THE UNITED STATES AND THE CONGO SUBMARINE VALLEY. A. C. Veatch and Paul A. Smith. 1939. 101 p., 5 charts, 10 pls., 28 figs. $4.80.
11. GEOLOGY AND PALEONTOLOGY OF THE MINGAN ISLANDS, QUEBEC. W. H. Twenhofel. 1938. 132 p., 24 pls., 1 fig., $1.50.
12. TERTIARY FLORA FROM THE RIO PICHILEUFU, ARGENTINA. E. W. Berry. 1938. 149 p., 56 pls., 11 figs. $2.10.
21. PALEONTOLOGY AND STRATIGRAPHY OF THE SILURIAN ROCKS OF THE PORT DANIEL BLACK CAPE REGION, GASPE. S. A. Northrop. 1939. 302 p., 28 pls., 1 fig. $3.00.
28. REVIEW OF THE PELYCOSAURIA. A. S. Romer and L. I. Price. 1949. 538 p., 46 pls., 71 figs. $4.20.
29. PHACOPID TRILOBITES OF NORTH AMERICA. D. M. Delo. 1940. 135 p., 13 pls., $1.20.
33. NUMERICAL STRUCTURE FACTOR TABLES. M. J. Buerger. Reprinted 1953. 119 p., 4 figs. (Cloth bound) $1.50.
37. OZARKIAN AND CANADIAN CEPHALOPODS, Part I, Nautilicones. E. O. Ulrich, Aug. F. Foerste, A. K. Miller and W. M. Furnish. 1942. 157 p., 57 pls., 23 figs. $2.40.
40. HADROSAURIAN DINOSAURS OF NORTH AMERICA. R. S. Lull and N. E. Wright. 1942. 242 p., 31 pls., 90 figs. $2.70.
49. OZARKIAN AND CANADIAN CEPHALOPODS. Part II: Brevicones. E. O. Ulrich, A. F. Foerste, and A. K. Miller. 1943. 240 p., 70 pls., 15 figs. $3.00.
60. GLACIAL MAP OF NORTH AMERICA. By Committee on Glacial Map. R. F. Flint, *Chairman*. 1945. $2.00.
62. CRUST OF THE EARTH. Arie Poldervaart, *Editor*. 1955. 762 p. $6.50.
63. DE NATURA FOSSILIUM, by Georgius Agricola. Translated by M. C. Bandy and J. A. Bandy. 1955. 240 p. $3.00.
64. MARYSVALE, UTAH, URANIUM AREA. Paul F. Kerr *et al.* 1957. 212 p., 21 pls., 106 figs. $6.00.

(*Continued on inside back cover*)

1969
UM

73° 00' 72° 30' 72° 00' 71° 30'

45° 30' 45° 00' 44° 30' 44°

Drummondville St. Lawrence River Nicolet River Danville Asbestos Shipton Pinnacle Weedon Richmond Marbleton Dudswell East Angus Sherbrooke Eaton River St. Francis River Acton Vale Yamaska River Black River Granby Shefford Mtn Waterloo Mt. Orford Eastman Brome Mtn Brome Lake Cowansville Farnham Yamaska Iberville Mt. Johnson St. Jean St. Bruno Mtn St. Hilaire Mtn Rougemont Yamaska Mtn St. Dominique Richelieu River Missisquoi Bay Bedford Sutton Mansonville Rosenberg Philipsburg Pinnacle Mtn Highgate Springs Swanton Enosburg Falls Richford Missisquoi River Bakersfield St. Albans Cambridge Lamoille River Morrisville Winooski River Winooski Burlington Lake Champlain Hardwick Montpelier Newport Lake Memphremagog Magog Coaticook River Coaticook Lake Massawippi Stoke Mountain Island Pond Willoughby Lake Barton River Black River Lyndonville Monadnock Mtn Goback Mtn St. Johnsbury Passumpsic River Connecticut River Lancaster Littleton Monroe

BAS DE STE. ROSE FAULT · FORTIERVILLE SYNCLINE · ST. BARNABÉ FAULT · "LOGANS LINE" T(?) · ST. DOMINIQUE THRUST · CHAMBLY-RICHELIEU FORELAND · DELSON FAULT · TRACY BROOK FAULT · HIGHGATE SPRINGS-ST. DOMINIQUE THRUST · PHILIPSBURG THRUST · ST. ALBANS-RICHFORD AXIAL SYNCLINORIUM · ENOSBURG MOUNTAIN · HIGHGATE SPRINGS THRUST · CHAMPLAIN THRUST · HINESBURG SYNCLINORIUM · CAMBRIDGE GREEN MOUNTAIN ANTICLINE · WORCESTER MTN ANTICLINE · LOWELL MTN ANTICLINE · MEMPHREMAGOG SYNCLINES · FALLS SYNCLINE · ANTICLINES · STOKE MOUNTAIN SYNCLINORIUM · CONNECTICUT VALLEY — GASPE ARCH · "MONROE LINE" · GARDNER MOUNTAIN ANTICLINE · AMMONOOSUC THRUST · SUTTON MOUNTAINS · QUEBEC · VERMONT · NEW HAMPSHIRE · NY · VT · NH · QUE

O(?) €um €l-€(?) D-S O €l-€(?) €um(?)

T(?) T

A — B

A' — B'

EXPLANATION

D-S — Devonian and Silurian rocks

Lake Memphremagog synclines: Peasley Pond conglomerate, Glenbrooke shale (slate), and Sargent Bay limestone.

Stoke Mountain anticline: Lake Aylmer group (conglomerate, calcareous shale, and limestone).

Connecticut Valley-Gaspe synclinorium: St. Francis group=Compton formation=Tomifobia slate (and limestone), in Quebec; Shaw Mountain formation (schistose conglomerate, quartzite, and limestone), Northfield slate, Ayers Cliff formation (arenaceous limestone to calcareous sandstone), Barton River formation=Waits River formation (interbedded slate and limestone), Standing Pond volcanics, and Westmore formation (schist, micaceous quartzite, and minor limestone), in Vermont; New Hampshire plutonic series(?).

Boundary Mountain anticlinorium: Meetinghouse slate and Gile Mountain (=Westmore) formation (schist, micaceous quartzite, and minor limestone) on the west limb, in Vermont; Clough quartzite (and conglomerate), Fitch formation (limestone and slate), and Littleton formation (slate, phyllite, schist, gneiss, greenstone, and amphibolite) on the east limb, in New Hampshire; Oliverian and New Hampshire plutonic series

Lower, Middle, and Upper Silurian and Lower Devonian — SILURIAN AND DEVONIAN

O — Ordovician rocks

Champlain-Richelieu foreland: Lacolle conglomerate and Stony Point, Iberville, Nicolet River, Pontgravé River, and Bécancour River formations (chiefly shales), in Quebec; Shelburne limestone (=Shelburne marble), Cutting dolomite, Bascom formation (limestone, dolomite, and quartzite), Bridport dolomite=Providence Island dolomite, the Day Point, Crown Point, Valcour, Isle La Motte, and Glens Falls limestones, the Cumberland Head and Stony Point formations (argillaceous limestone and calcareous shale), Iberville formation (shale and siltstone), and Hathaway formation (argillite and chert), in Vermont.

Highgate Springs-St. Dominique slice: Beldens formation (limestone and dolomite), Chazy limestone=St. Dominique limestone, Black River and Trenton limestones, and Utica and Lorraine shales, in Quebec; Beldens formation (limestone and dolomite), Cerman quartzite, Youngman formation (argillitic limestone and argillite), Isle La Motte and Glens Falls limestones, and Hortonville slate=Canajoharie slate, in Vermont.

Phillipsburg slice: Strites Pond and Wallace Creek limestones, Morgan Corner dolomite, Hastings Creek, Naylor Ledge, and Lake Hill limestones, in Vermont and Quebec; Solomons Corner, St. Armand, and Coray limestones, Basswood Creek shale, Mystic conglomerate, and Stanbridge slate, in Quebec.

St. Albans synclinorium: Highgate formation (limestone and slate) and Moreas Line slate=Stanbridge slate, in Vermont and Quebec. Also, in Quebec, north of the Yamaska River, successively(?) the upper part of the Sillery formation (red slate and green graywacke), limestone near Acton Vale, and the Farnham formation (gray to black shale and limestone). Units included in the Hinesburg synclinorium in Vermont, but not shown on the correlation table in this belt (pl. 3), are the Shelburne marble, Cutting dolomite, and Bascom formation (dolomite and limestone) (Cady, 1945, pl. 10).

Enosburg Falls anticline: the upper part of the "Sillery" formation="Caldwell facies" (massive green graywacke and granule conglomerate) in Quebec, west and northwest of Richmond.

Green Mountain axial anticline (east limb): All but lowermost parts of the upper Mansonville formation="Caldwell facies" (phyllitic graywacke, granule conglomerate, greenstone, and purple to red slate, in Quebec, and the Stowe formation (quartz-sericite-chlorite schist, phyllitic graywacke, and greenstone), in Vermont.

Lake Memphremagog synclines and Lowell Mountain and Stoke Mountain anticlines: "Caldwell" formation (graywacke, slate, siltstone, and quartzite), Beauceville formation (slate and quartzite), and Bolton volcanics, in Quebec; Stowe formation (quartz-sericite-chlorite schist) and Moretown formation (granulite, quartzite, and slate), in Vermont. Units included in the Worcester Mountain anticline in Vermont are the same as those in the Lowell Mountain anticline.

Boundary Mountain anticlinorium: Albee formation (phyllite, slate, quartzite), Ammonoosuc volcanics, and Partridge formation (chiefly slate) in New Hampshire; Albee and Ammonoosuc in Vermont; Highlandcroft plutonic series

Lower and Middle Ordovician — ORDOVICIAN

€um — Upper and Middle Cambrian rocks

Champlain-Richelieu foreland: Clarendon Springs dolomite.

Philipsburg slice: Rock River dolomite in Quebec and Vermont.

St. Albans synclinorium: Saxe Brook dolomite and Gorge formation (dolomite, slate, and limestone), in Quebec and nearby areas in Vermont, and also in Quebec north of the Yamaska River, the middle part of the Sillery formation (gray sandstones(?); in Vermont, the Rugg Brook dolomite (also conglomerate and sandstone), St. Albans slate, Mill River conglomerate, Skeels Corners slate, Rockledge conglomerate, Saxe Brook dolomite, Hungerford slate, and Gorge formation. Units included in the Hinesburg synclinorium, but not shown on the correlation table in this belt (pl. 3), are the Danby formation (dolomite and quartzite) and the Clarendon Springs dolomite (Cady, 1945, pl. 10).

Enosburg Falls anticline: Scottsmore quartzite (and conglomerate) and Sweetsburg slate, in Quebec, and Rugg Brook dolomite (also conglomerate and quartzite), Skeels Corners slate, Rockledge conglomerate, and Hungerford slate, in Vermont.

Cambridge syncline: Sweetsburg slate and Melbourne limestone in Quebec; possibly the Scottsmore quartzite and Sweetsburg slate (and limestone), in Vermont.

Green Mountain axial anticline: Lower part of the Mansonville formation (gray quartzite, dark phyllite, and granule conglomerate) in Quebec; Ottauquechee formation (carbonaceous phyllite and quartzite, granule conglomerate) and lowermost Stowe formation (quartz-sericite-chlorite phyllite and granule conglomerate) in Vermont

Middle and Upper Cambrian — CAMBRIAN

€l-€(?) — Lower Cambrian and Cambrian(?) rocks

St. Albans synclinorium: Dunham dolomite in Quebec and Vermont succeeded by Parker slate in Vermont. Units included in the Hinesburg synclinorium, but not shown on the correlation table in this belt (pl. 3), are the Monkton quartzite (and dolomite), and the Winooski dolomite (Cady, 1945, pl. 10).

Enosburg Falls anticline: Tibbit Hill volcanics, Cell Mill slate, Pinnacle formation (graywacke, phyllite, arkose), White Brook dolomite, West Sutton slate, Gilman quartzite (also phyllite), Dunham dolomite, and Oak Hill=Parker slate.

Cambridge syncline: "Bonsecours" formation (quartz-sericite-chlorite schist) of Osberg in Quebec and "Jay Peak series" (chloritic, micaceous, quartzitic, and carbonaceous schist) of Booth in Vermont.

Green Mountain axial anticline: Sutton schist (carbonaceous and noncarbonaceous quartz-sericite-chlorite schist) in Quebec and the lithologically similar Camels Hump group, including the Belvidere Mountain amphibolite, in Vermont

Cambrian(?) and Lower Cambrian — CAMBRIAN(?) AND CAMBRIAN

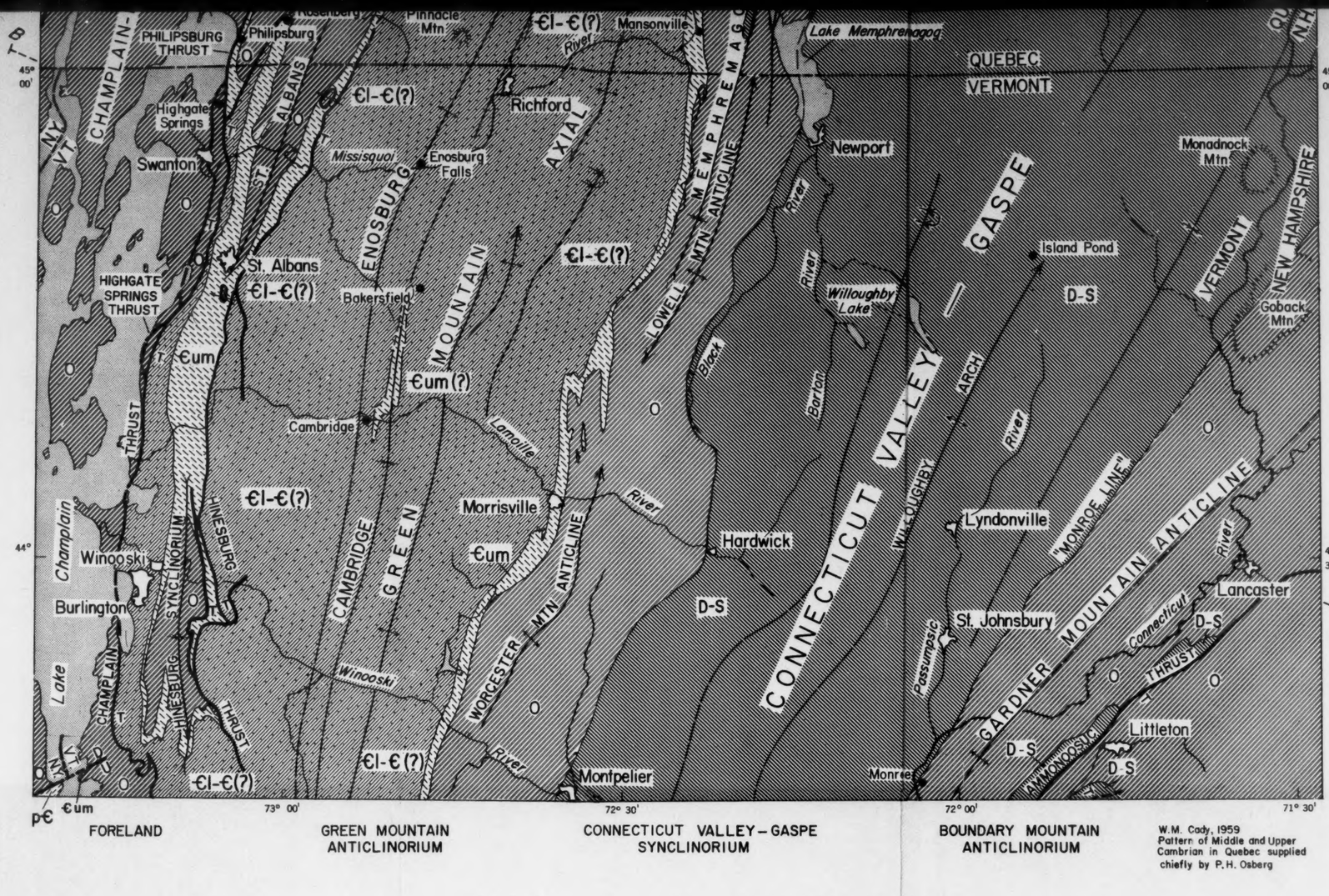
pC — Precambrian rocks

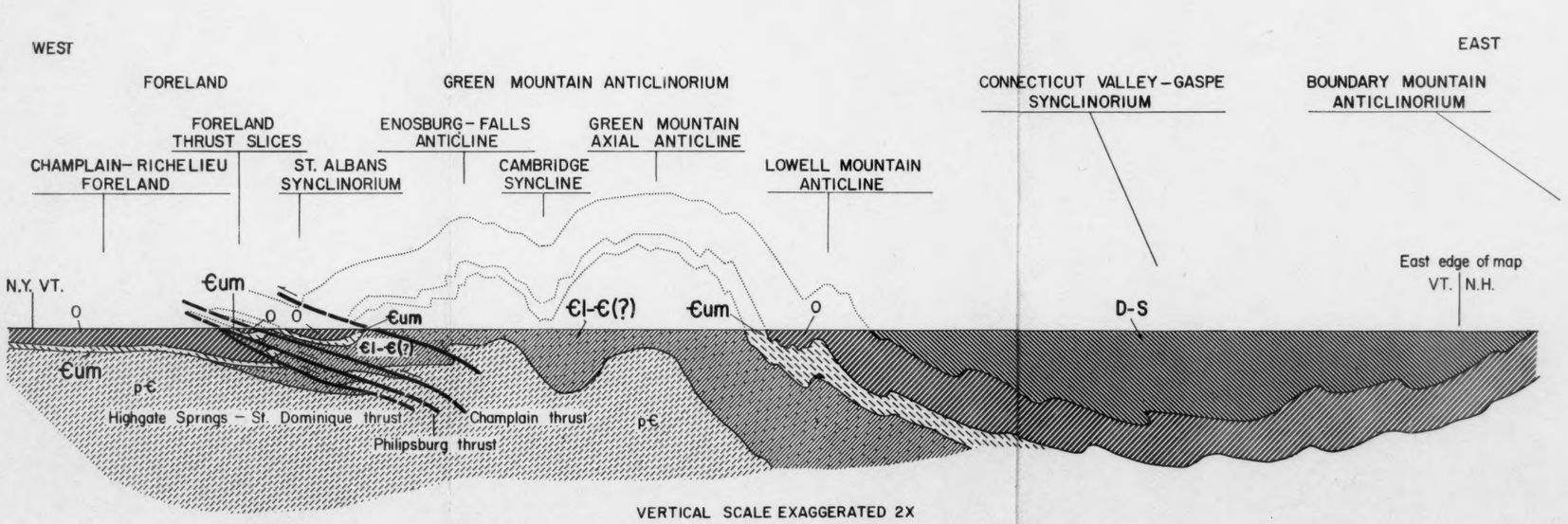
Champlain-Richelieu foreland: Gneiss of the Grenville series, also metagabbro and granite, of the Adirondack massif on the west shore of Lake Champlain in New York (not included in the present discussion)

PRECAMBRIAN

Contact

U / D — Normal fault





GENERALIZED STRUCTURE SECTION AT INTERNATIONAL BOUNDARY

volcanics, and Partridge formation (chiefly slate) in New Hampshire; Albee and Ammonoosuc in Vermont; Highlandcraft plutonic series

€um — Upper and Middle Cambrian rocks

Champlain-Richelieu foreland: Clarendon Springs dolomite.
Philipsburg slice: Rock River dolomite in Quebec and Vermont.
St. Albans synclinorium: Saxe Brook dolomite and Gorge formation (dolomite, slate, and limestone), in Quebec and nearby areas in Vermont, and also in Quebec north of the Yamaska River, the middle part of the Sillery formation (gray sandstones(?)); in Vermont, the Rugg Brook dolomite (also conglomerate and sandstone), St. Albans slate, Mill River conglomerate, Skeels Corners slate, Rockledge conglomerate, Saxe Brook dolomite, Hungerford slate, and Gorge formation. Units included in the Hinesburg synclinorium, but not shown on the correlation table in this belt (pl. 3), are the Danby formation (dolomite and quartzite) and the Clarendon Springs dolomite (Cady, 1945, pl. 10).
Enosburg Falls anticline: Scottsmore quartzite (and conglomerate) and Sweetsburg slate, in Quebec, and Rugg Brook dolomite (also conglomerate and quartzite), Skeels Corners slate, Rockledge conglomerate, and Hungerford slate, in Vermont.
Cambridge syncline: Sweetsburg slate and Melbourne limestone in Quebec; possibly the Scottsmore quartzite and Sweetsburg slate (and limestone), in Vermont.
Green Mountain axial anticline: Lower part of the Hansonville formation (gray quartzite, dark phyllite, and granule conglomerate) in Quebec; Ottauquechee formation (carbonaceous phyllite and quartzite, granule conglomerate) and lowermost Stowe formation (quartz-sericite-chlorite phyllite and granule conglomerate) in Vermont

Middle and Upper Cambrian — CAMBRIAN

€l-€(?) — Lower Cambrian and Cambrian(?) rocks

St. Albans synclinorium: Dunham dolomite in Quebec and Vermont succeeded by Parker slate in Vermont. Units included in the Hinesburg synclinorium, but not shown on the correlation table in this belt (pl. 3), are the Monkton quartzite (and dolomite), and the Winooski dolomite (Cady, 1945, pl. 10).
Enosburg Falls anticline: Tibbit Hill volcanics, Call Mill slate, Pinnacle formation (graywacke, phyllite, arkose), White Brook dolomite, West Sutton slate, Gilman quartzite (also phyllite), Dunham dolomite, and Oak Hill-Parker slate.
Cambridge syncline: "Bonsecours" formation (quartz-sericite-chlorite schist) of Osberg in Quebec and "Jay Peak series" (chloritic, micaceous, quartzitic, and carbonaceous schist) of Booth in Vermont.
Green Mountain axial anticline: Sutton schist (carbonaceous and noncarbonaceous quartz-sericite-chlorite schist) in Quebec and the lithologically similar Camels Hump group, including the Belvidere Mountain amphibolite, in Vermont

Cambrian(?) and Lower Cambrian — CAMBRIAN(?) AND CAMBRIAN

pC — Precambrian rocks

Champlain-Richelieu foreland: Gneiss of the Grenville series, also metagabbro and granite, of the Adirondack massif on the west shore of Lake Champlain in New York (not included in the present discussion)

PRECAMBRIAN

Contact

U / D — Normal fault

Thrust fault
T on side of upper plate

Anticline
Showing direction of plunge

Syncline
Showing direction of plunge

Overturned syncline
Showing direction of plunge

Structural features explained above are dashed where inferred

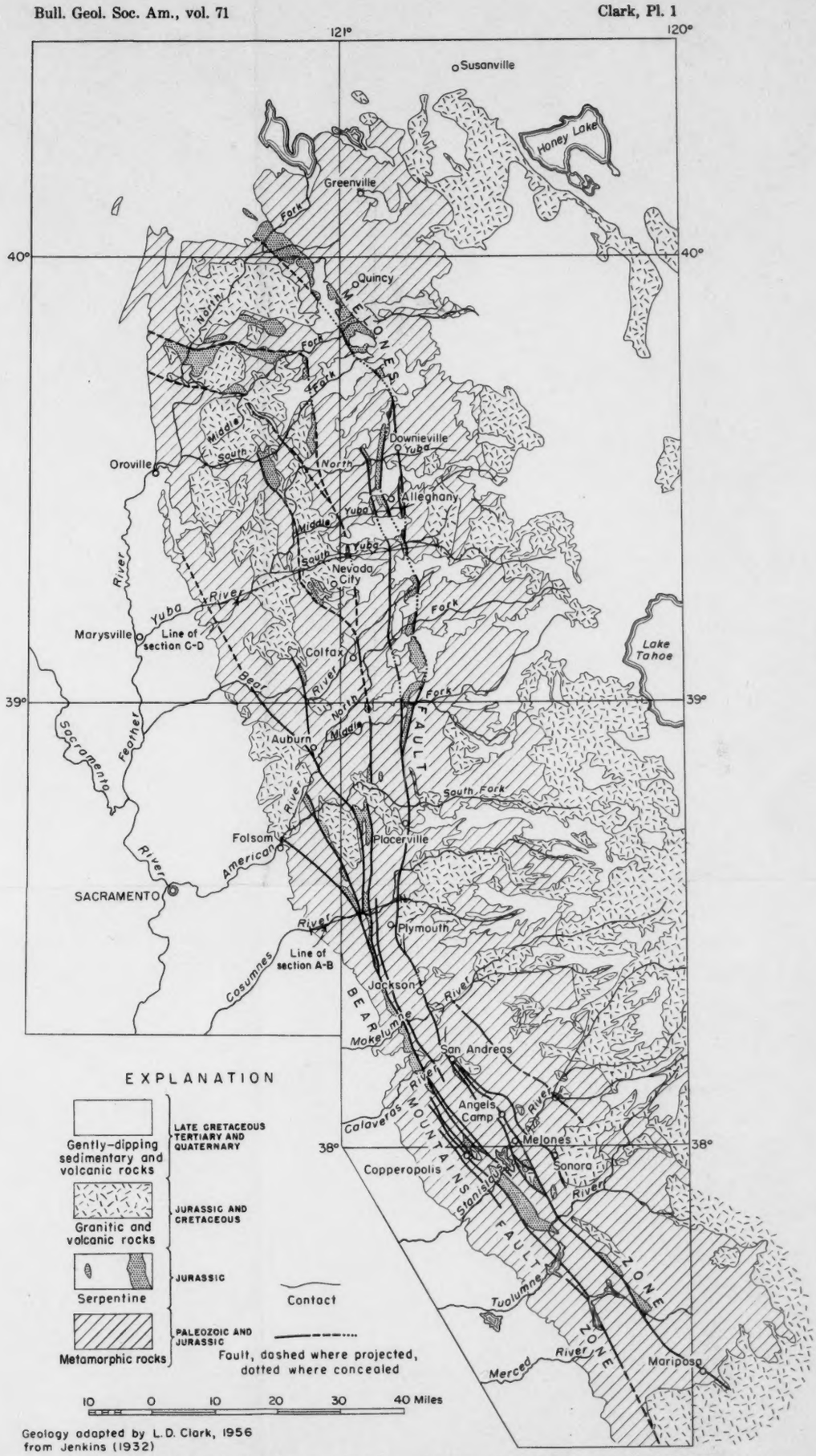
Direction of overthrust movement in klippe

Isolated mountain groups
Underlain chiefly by granitic rocks of the Monteregian and White Mountain plutonic series

Mountain summits

TECTONIC MAP AND SECTION OF NORTHERN VERMONT AND NEARBY PARTS OF SOUTHERN QUEBEC

See Plate 2 for restored sections A-A′ and B-B′.



Geology adapted by L.D. Clark, 1956 from Jenkins (1932)

OUTLINE GEOLOGIC MAP OF THE WESTERN SIERRA NEVADA, CALIFORNIA, SHOWING FAULTS OF THE FOOTHILLS FAULT SYSTEM

LEGEND

WESTERN AREA EASTERN AREA

QUATERNARY — Ql — LANDSLIDES

MIO-PLIOCENE — Y — YAKIMA-BASALT

Sq — SNOQUALMIE GRANODIORITE

Ry — POST-ROSLYN RHYOLITE (MAY LOCALLY INCLUDE OLDER RHYOLITE)

H — HOWSON ANDESITE

APPROXIMATE MEAN DECLINATION 1902

1 0 1 2 3 4 5 6 7 8 9 Miles

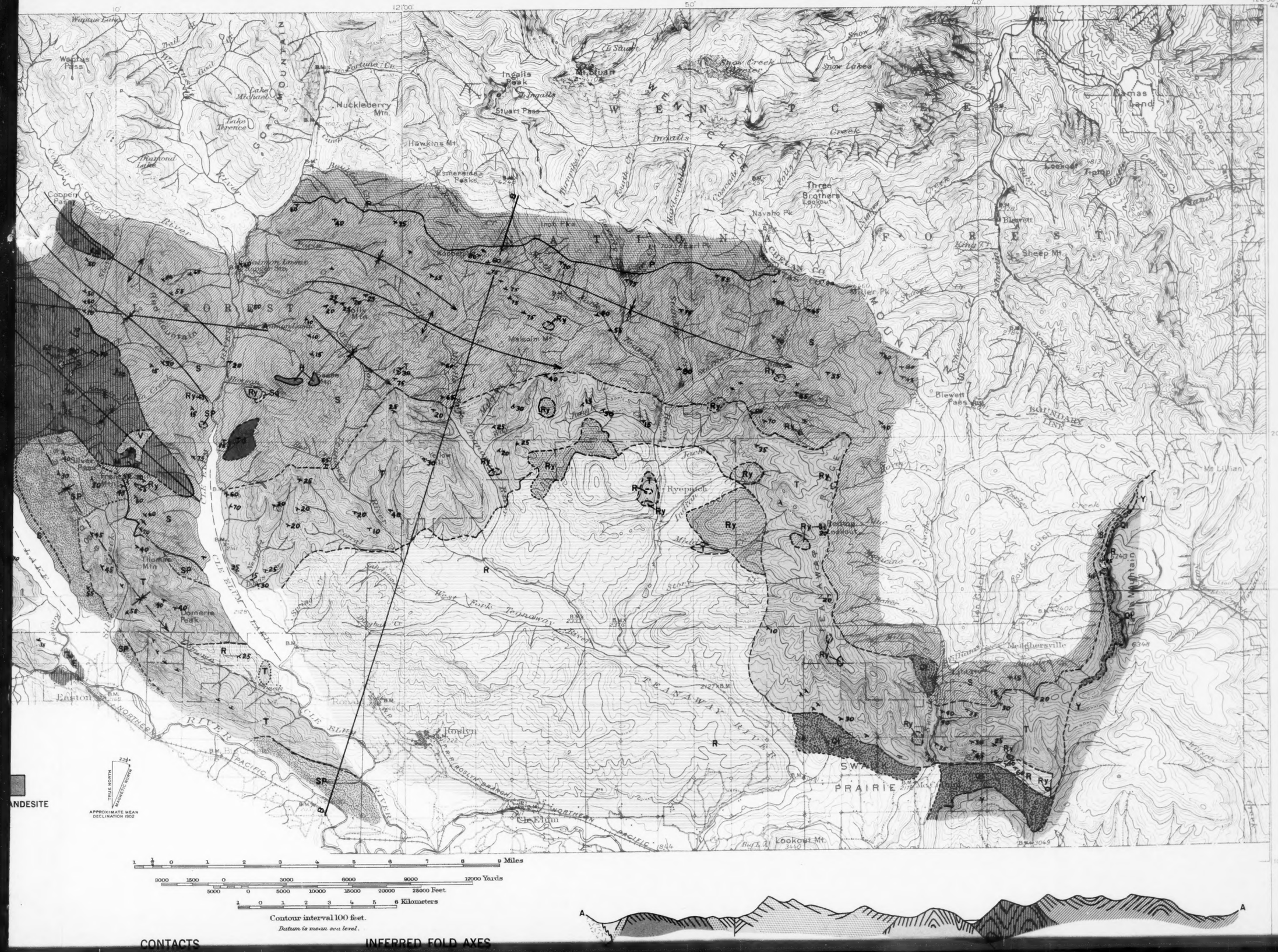
3000 1500 0 3000 6000 9000 12000 Yards

5000 0 5000 10000 15000 20000 25000 Feet

1 0 1 2 3 4 5 6 Kilometers

Contour interval 100 feet.

Datum is mean sea level.


Wenatchee Forest
Snoqualmie Forest
Cle Elum River
Yakima River
Teanaway River
Swauk Prairie
Ingalls Peak
Stuart Pass
Nuckleberry Mtn.
Hawkins Mt.
Esmeralda Peaks
Three Brothers Lookout
Navaho Pk.
Miller Pk.
Sheep Mt.
Blewett
Blewett Pass
Lemas Land
Lookout
Malsom Mt.
Ryepatch
Roslyn
Meighersville
Kookout Mt.
Mt. Lillian
Thomas Mtn.
Domerie Peak
Easton
Cle Elum
Boundary Line
Andesite
Contour interval 100 feet.
Datum is mean sea level.
CONTACTS
INFERRED FOLD AXES

LEGEND

WESTERN AREA EASTERN AREA

QUATERNARY — Qls — LANDSLIDES

MIO-PLIOCENE — Y — YAKIMA-BASALT

Sq — SNOQUALMIE GRANODIORITE

Ry — POST-ROSLYN RHYOLITE (MAY LOCALLY INCLUDE OLDER RHYOLITE)

OLIGOCENE? — K — KEECHELUS ANDESITE

EOCENE — R — ROSLYN ARKOSE

N — NACHES FM. (WITH INCLUDED RHYOLITES)

T — TEANAWAY BASALT

Td — DIABASE

EOCENE? — MC — MT. CATHERINE RHYOLITE

SP — SILVER PASS VOLCANIC ROCKS

G — GUYE FM.

PALEOCENE? — S — SWAUK FM.

H — HOWSON ANDESITE

PRE-TERTIARY — DENNY FM.

PRE-TERTIARY — E — EASTON SCHIST

P — PERIDOTITE AND OLDER ROCKS

M — MARBLE (MINOR AREAS WEST OF CLE ELUM LAKE)

V — VOLCANIC ROCKS

APPROXIMATE MEAN DECLINATION 1902

1 ½ 0 1 2 3 4 5 6 7 8 9 Miles

3000 1500 0 3000 6000 9000 12000 Yards

5000 0 5000 10000 15000 20000 25000 Feet

1 0 1 2 3 4 5 6 Kilometers

Contour interval 100 feet.

Datum is mean sea level.

CONTACTS

LOCATED

INFERRED

FAULT

INFERRED FOLD AXES

SYNCLINE (SHOWING DIRECTION OF PITCH)

ANTICLINE

STRIKE & DIP OF BEDS

×30

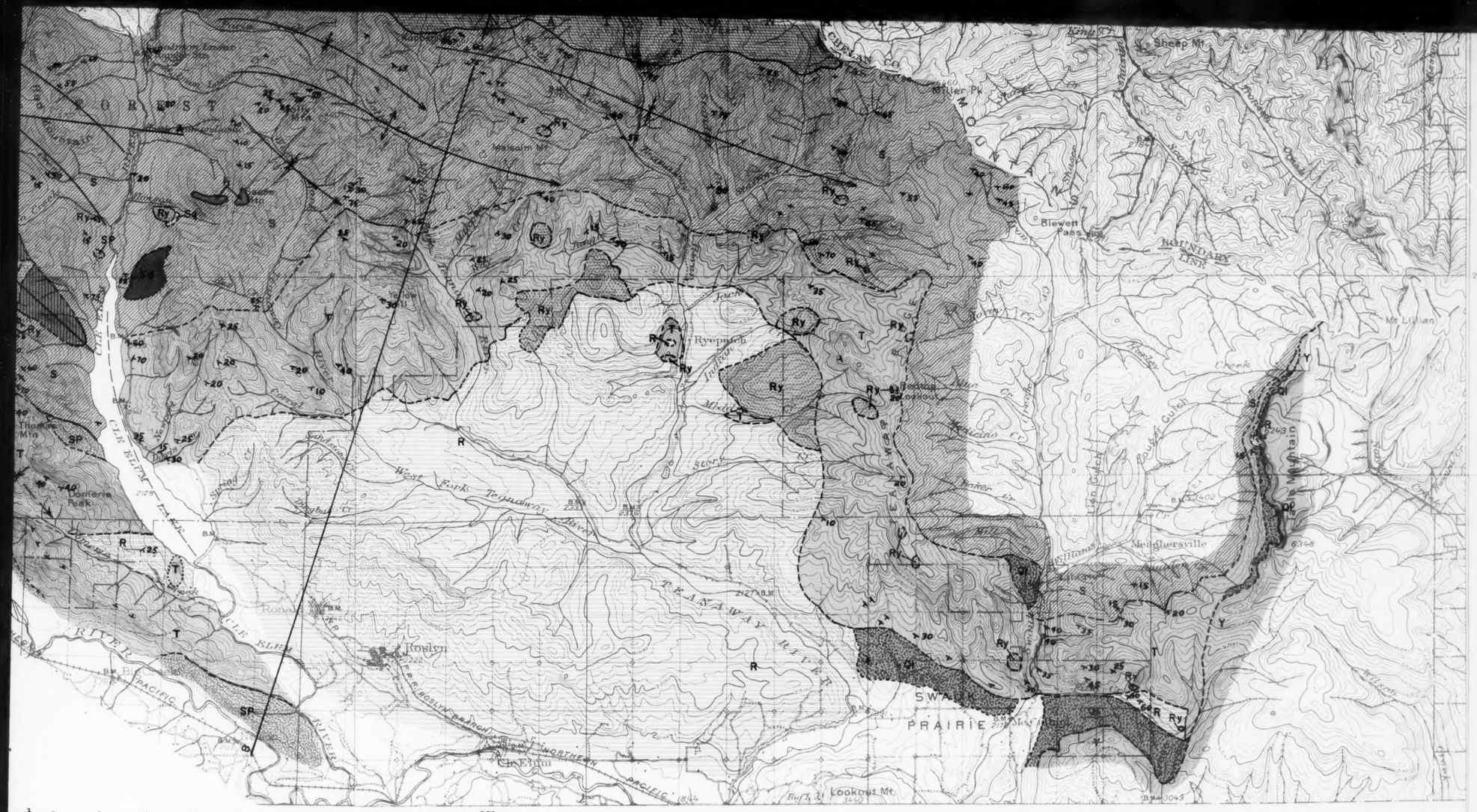
ATTITUDE OF FOLATION

↳20

A

B

GEOLOGIC MAP AND CROSS SECTIONS, CENTRAL CASCADES, WASHINGTON



1 0 1 2 3 4 5 6 7 8 9 Miles

3000 1500 0 3000 6000 9000 12000 Yards

5000 0 5000 10000 15000 20000 25000 Feet

1 0 1 2 3 4 5 6 Kilometers

Contour interval 100 feet.

Datum is mean sea level.

CONTACTS

LOCATED

INFERRED

FAULT

V

VOLCANIC ROCKS
(F CLE ELUM LAKE)

INFERRED FOLD AXES

SYNCLINE
(SHOWING DIRECTION OF PITCH)

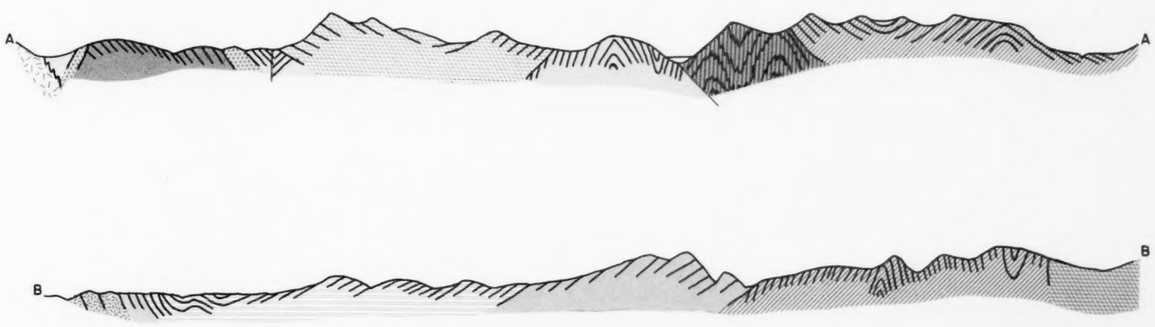
ANTICLINE

STRIKE & DIP OF BEDS

30

ATTITUDE OF FOLATION

20



Northern margin of Swauk arkose from U.S.G.S. Folios 106 & 109.

Base map: Snoqualmie Pass & Mt. Stuart quadrangles U.S.G.S.

Geology: R. J. Foster, 1956

120°45' 40' 50' 12700'

Williams & Heintz Map Corporation, Washington 27, D.C.

OLOGIC MAP AND CROSS SECTIONS, CENTRAL CASCADES, WASHINGTON

116°05'

45,000

40,000

40,000

35,000

35°
32½

35,000

NORTH ISLAND

MIDDLE ISLAND

TOKOPAH AND TIDE WATER (Abandoned)

RAILROAD VALLEY

BM 1101.8

ROAD CANYON

VAN FAULT

A

A'

J

J'

116°05'

35,000

30,000

25,000

40,000

35°
32½

35,000

B'

C'

D'

MEEKER FAULT

JEEP VALLEY

VAN FAULT

1500

1700

1900

2100

2300

2500

2700

2900

MIDDLE ISLAND

SOUTH ISLAND

RAILROAD VALLEY

ROAD CANYON

ROAD VALLEY

VAN FAULT

RIGGS FAULT

Riggs Mine

Rock Eagle Mine

Lone Wolf Mine

Brecciated Riggs formation

XBM 1101.8

30,000

45,000

25,000

J''

K

A

EXPLANATION

CENOZOIC

Alluvium, taluses, slump blocks, terrace gravels, fan gravels and fanglomerates

br

Breccia, composed dominantly of Goodsprings dolomite

g / v

Gravels, g Volcanics, v
Tertiary (?)

CENOZOIC

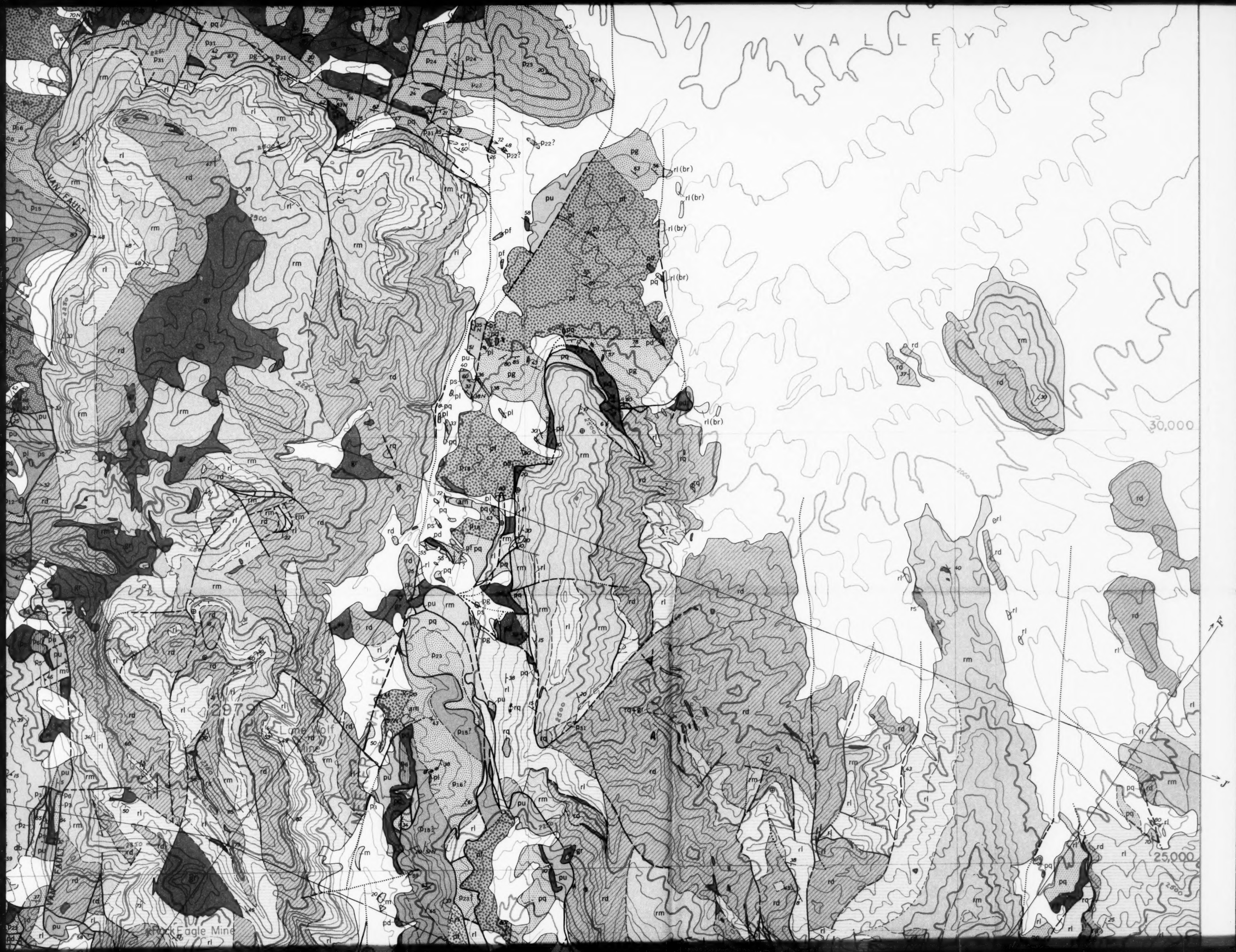
am / db

Amphibole-rich rock, probably sedimentary, am; intrusive rock, diabase and/or amphibolitized, db

PAHRUMP GROUP SUBDIVIDED BY THE DOMINANT LITHOLOGY

pu

Undifferentiated and brecciated Pahrump, some Riggs formation


VALLEY
VAN FAULT
Lone Wolf Mine
Rock Eagle Mine
30,000
25,000

EXPLANATION

MESOZOIC OR CENOZOIC — CENOZOIC

- Alluvium, taluses, slump blocks, terrace gravels, fan gravels and fanglomerates
- br — Breccia, composed dominantly of Goodsprings dolomite
- g / v — Gravels, g; Volcanics, v; Tertiary (?)
- t — Talc (age unknown)
- ap — Aplite
- gr — Granitic rock
- i — Inclusion in granitic rock (age unknown)

PALEOZOIC (?) — Riggs formation

- rq / rs — Quartzite, rq; and thin-bedded, clastic sedimentary rock, rs
- rl — Limestone (weathers bluish or light grey)
- rm — Mixed carbonates (often well bedded)
- rd — Dolomite (weathers buff)

PAHRUMP GROUP SUBDIVIDED INTO NUMBERED MEMBERS
(see text for details)

PRECAMBRIAN — Pahrump group

- am / db — Amphibole-rich rock, probably sedimentary, am; intrusive rock, diabase and/or amphibolitized, db

PAHRUMP GROUP SUBDIVIDED BY THE DOMINANT LITHOLOGY

- pu — Undifferentiated and brecciated Pahrump, some Riggs formation
- pf — Fanglomerate and conglomerate; cobbles of various rock types
- pc — Conglomerate with cobbles of nearly pure quartz
- pg — Granular rock, dominantly impure sandstone and quartzite
- pq — Quartzite, generally vitreous and massive
- ps — Sedimentary rock, clastic and well-bedded, dominantly siltstone
- pd — Dolomite (weathers orange-brown), generally interbedded with quartzite
- pl — Limestone (weathers white), sometimes interbedded with quartzite
- m — Pre-Pahrump granitic and metamorphic rock

Contact, long dashes where approximately located, short dashes where hypothetical

Contact, known fault; dashed where approximately located, dotted where hypothetical

Marker bed

Axis of syncline (minor)

A——A′ Line of section (F-F′ and J-J′ on Pl. 2)

58 Dip and strike of bedding

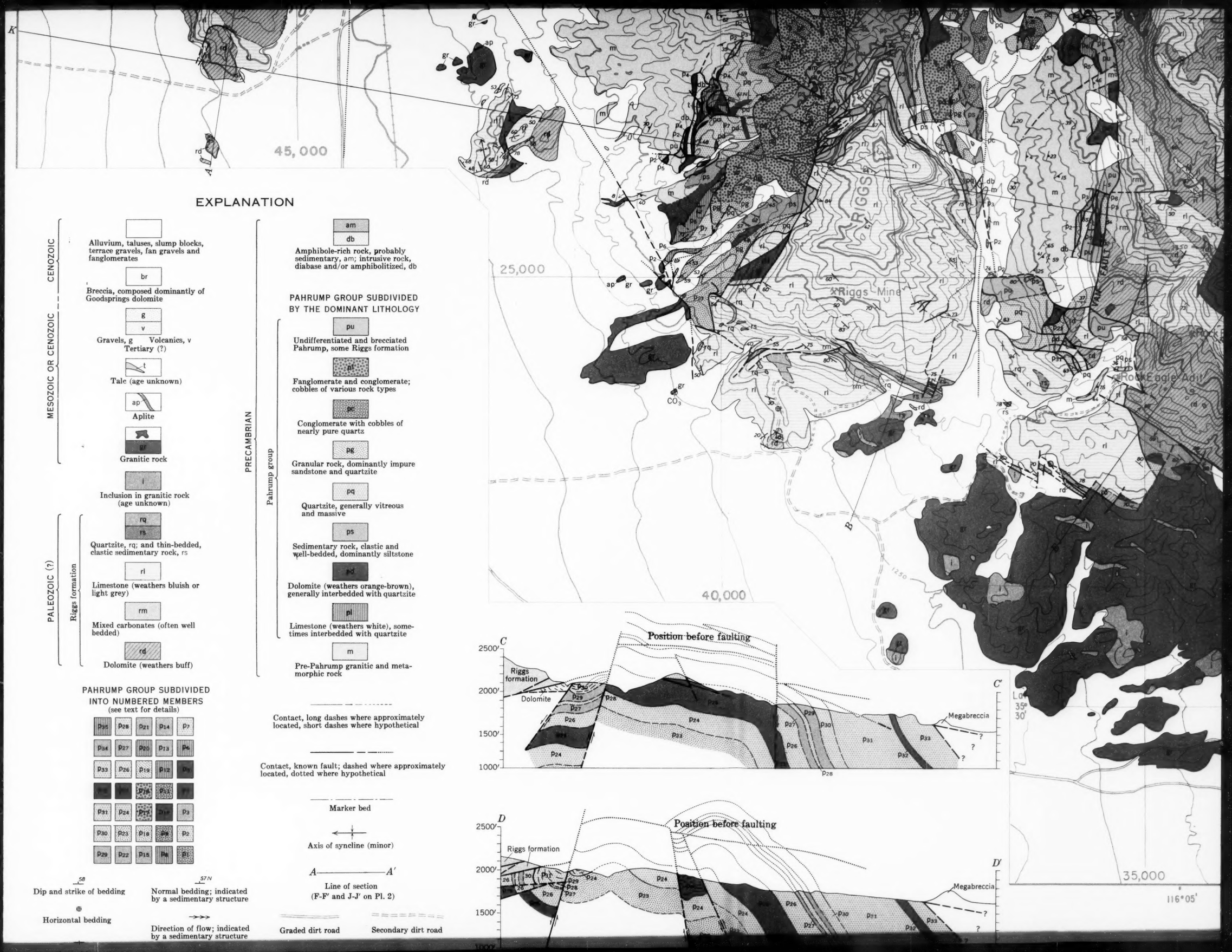
57 N Normal bedding; indicated by a sedimentary structure

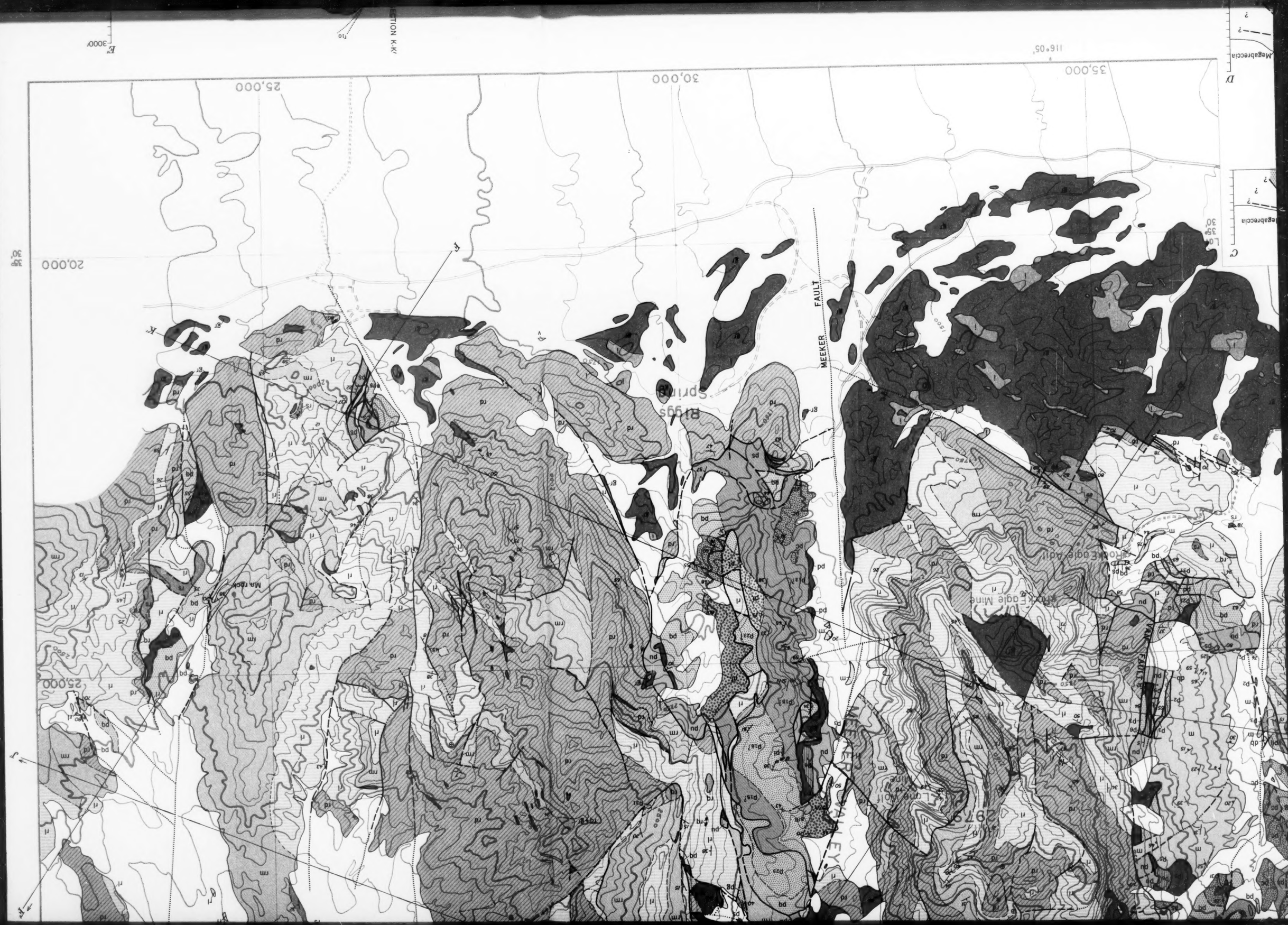
⊕ Horizontal bedding

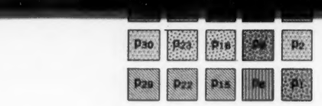
Direction of flow; indicated by a sedimentary structure

Graded dirt road

Secondary dirt road




MEEKER FAULT
Rhyolite Eagle Mine
Mt. Rock
Blyss Spring
20,000
25,000
30,000
35,000
116°05'
30'
37°
E
E'
F
F'
K
K'
C
C'
D
D'
Megabreccia
Megabreccia



Dip and strike of bedding

Horizontal bedding

Vertical beds

Overturning; indicated by a sedimentary structure

Normal bedding; indicated by a sedimentary structure

Direction of flow; indicated by a sedimentary structure

Dip and strike of foliation

Dip (true) of fault

Axis of syncline (minor)

A——————A′

Line of section
(F-F′ and J-J′ on Pl. 2)

Graded dirt road

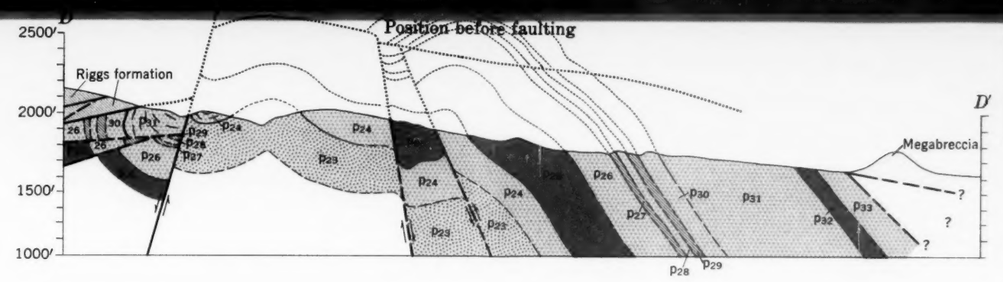
Truck tracks

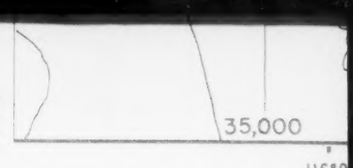
Mine

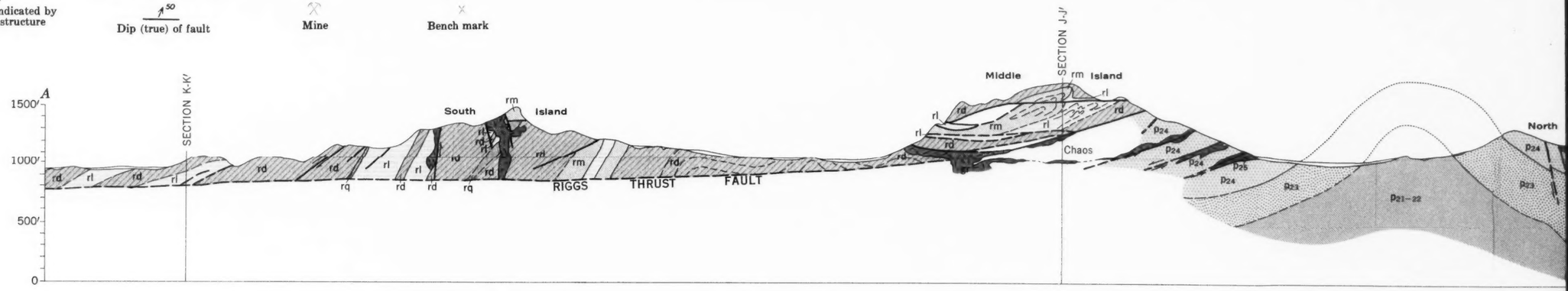
Secondary dirt road

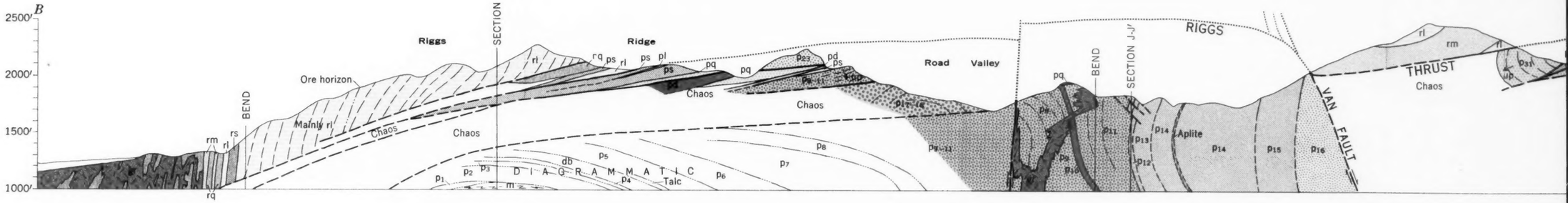
Railroad, abandoned

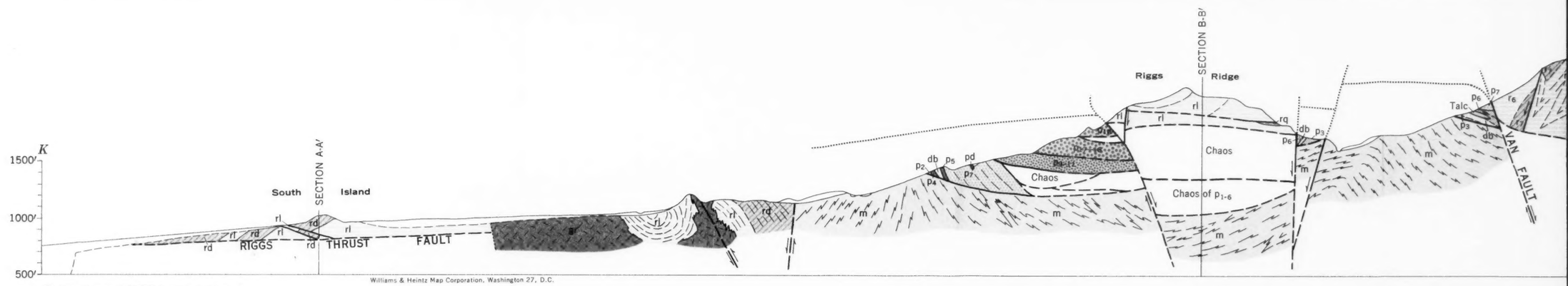
Bench mark











Williams & Heintz Map Corporation, Washington 27, D.C.

Horizontal control: USGS Avawatz quadrangle
Vertical control: Barometric observations

SILURIAN HILLS, SAN BERNARDI

GEOLOGIC MAP OF WESTERN PORTION, A

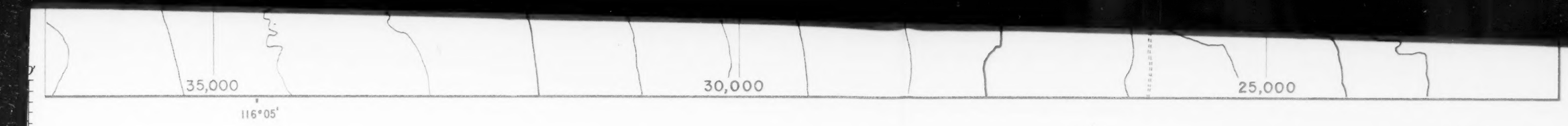
See plate 2 for complete sections F-F′

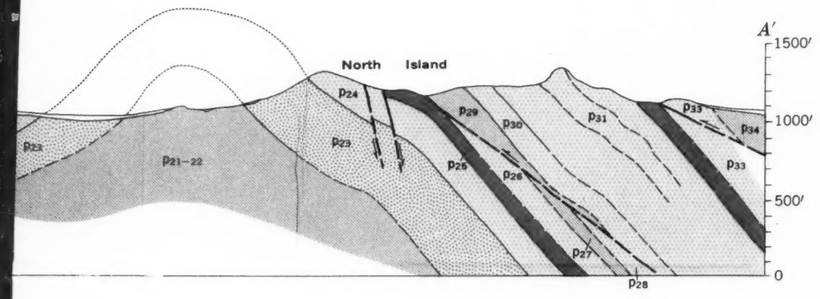
SCALE 1:9600

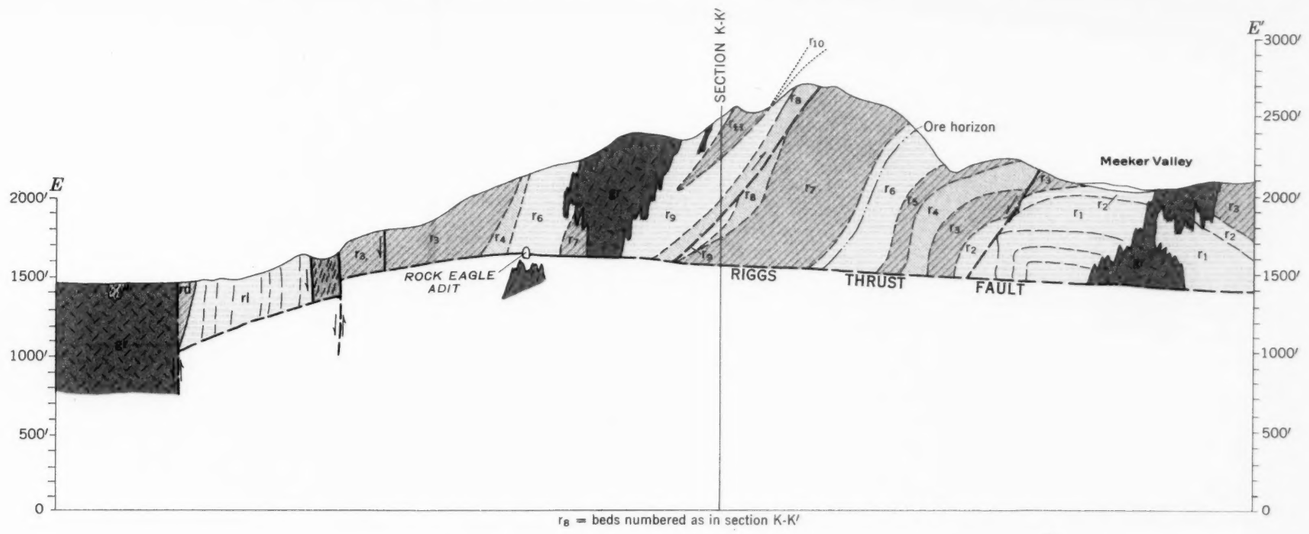
500 0 1000 200

MAGNETIC DECLINATION 15°45′ EAS
CONTOUR INTERVAL 50 FEET

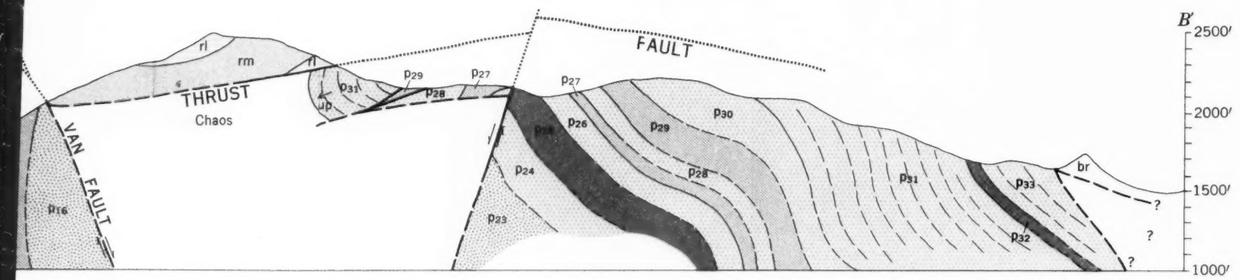
1951

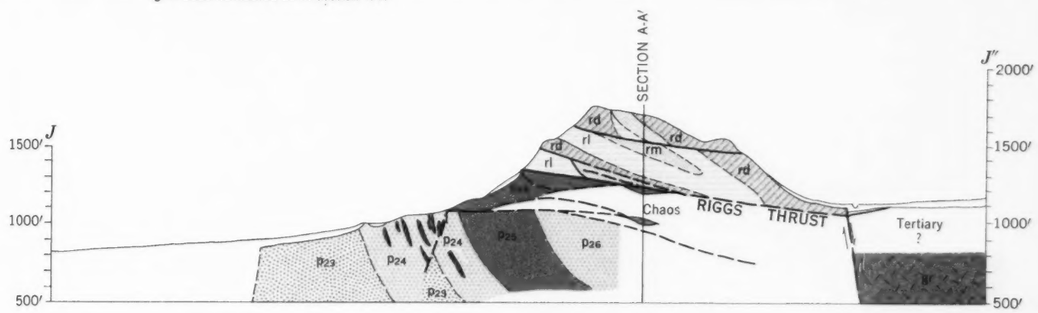


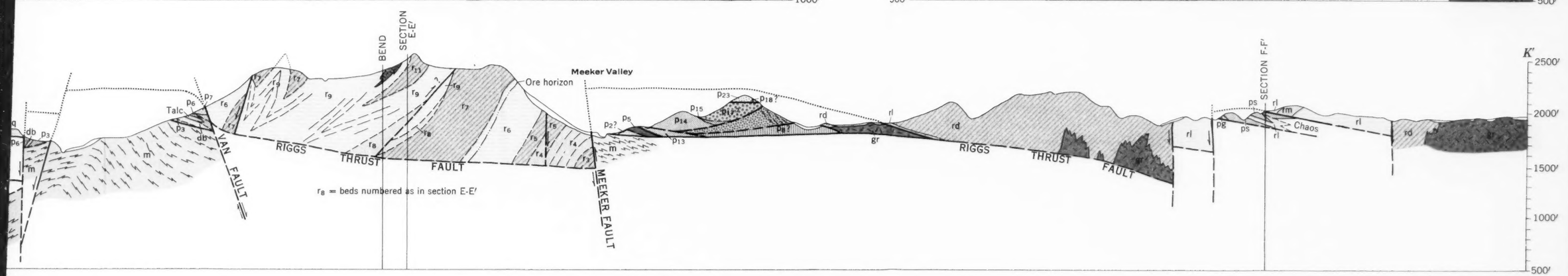




r8 = beds numbered as in section K-K'







r8 = beds numbered as in section E-E'

SAN BERNARDINO COUNTY, CALIFORNIA

OF WESTERN PORTION, AND STRUCTURE SECTIONS

See plate 2 for complete sections F-F' and J-J'

SCALE 1:9600

500 0 1000 2000 3000 FEET

MAGNETIC DECLINATION 15°45' EAST
CONTOUR INTERVAL 50 FEET

1951

Topography sketched from aerial photographs.
Geology mapped on aerial photographs 1948-50
by Donald H. Kupfer

LEGEND

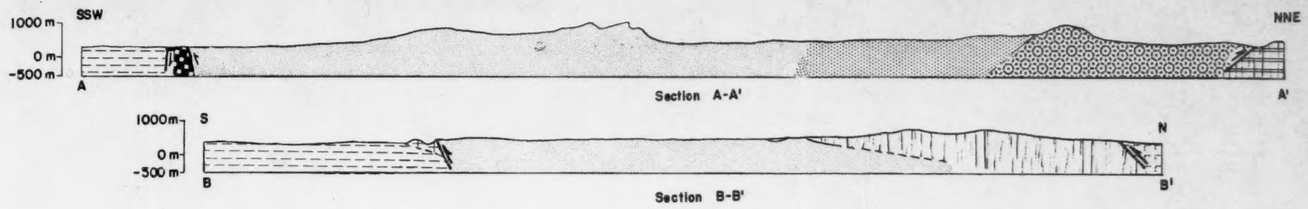
Upper Cretaceous to Paleocene

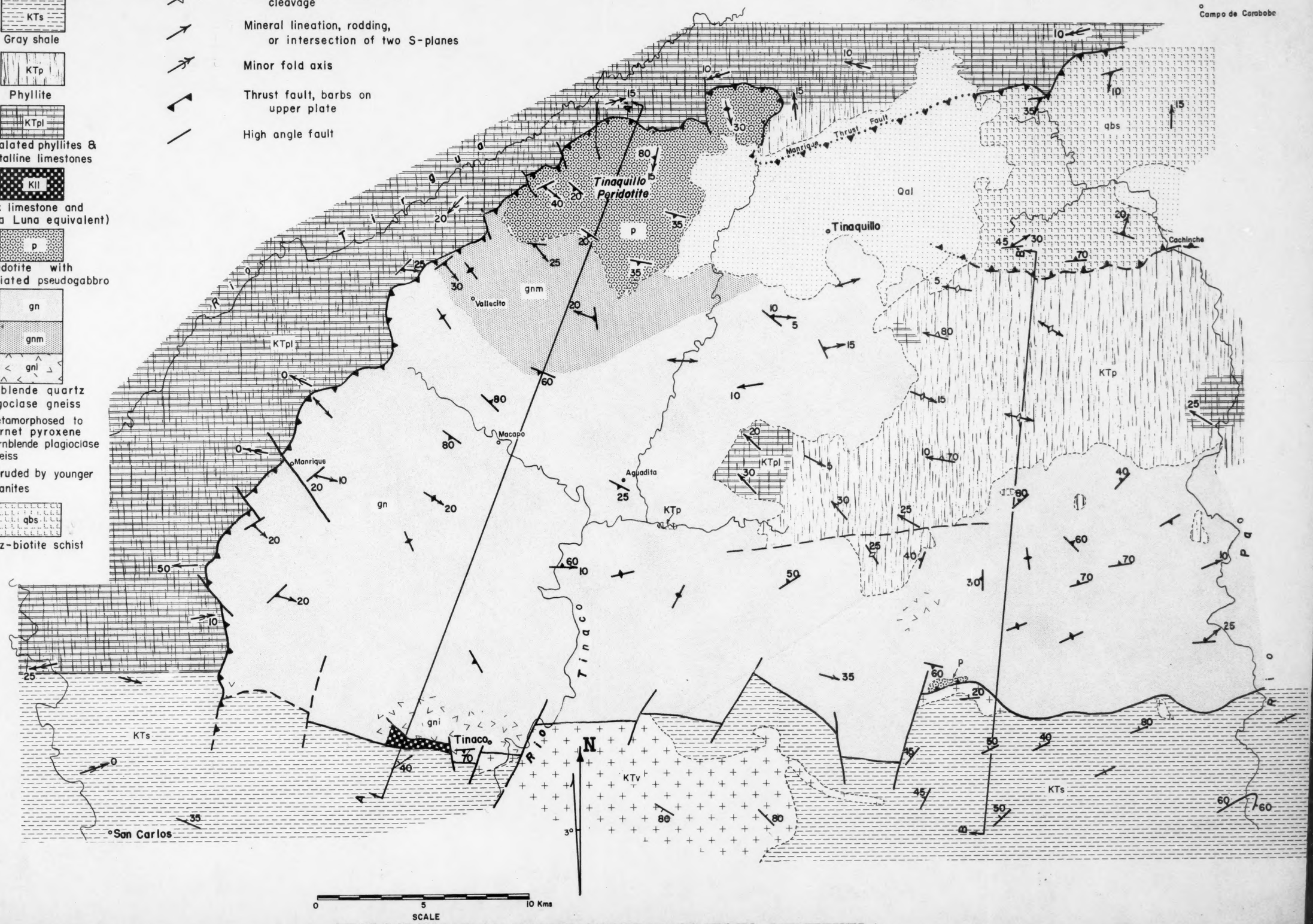
- Qal — Alluvium
- KTv — Basaltic & andesitic volcanics
- KTs — Gray shale
- KTp — Phyllite
- KTpl — Intercalated phyllites & crystalline limestones
- Kll — Black limestone and shale (La Luna equivalent)

Cretaceous ?

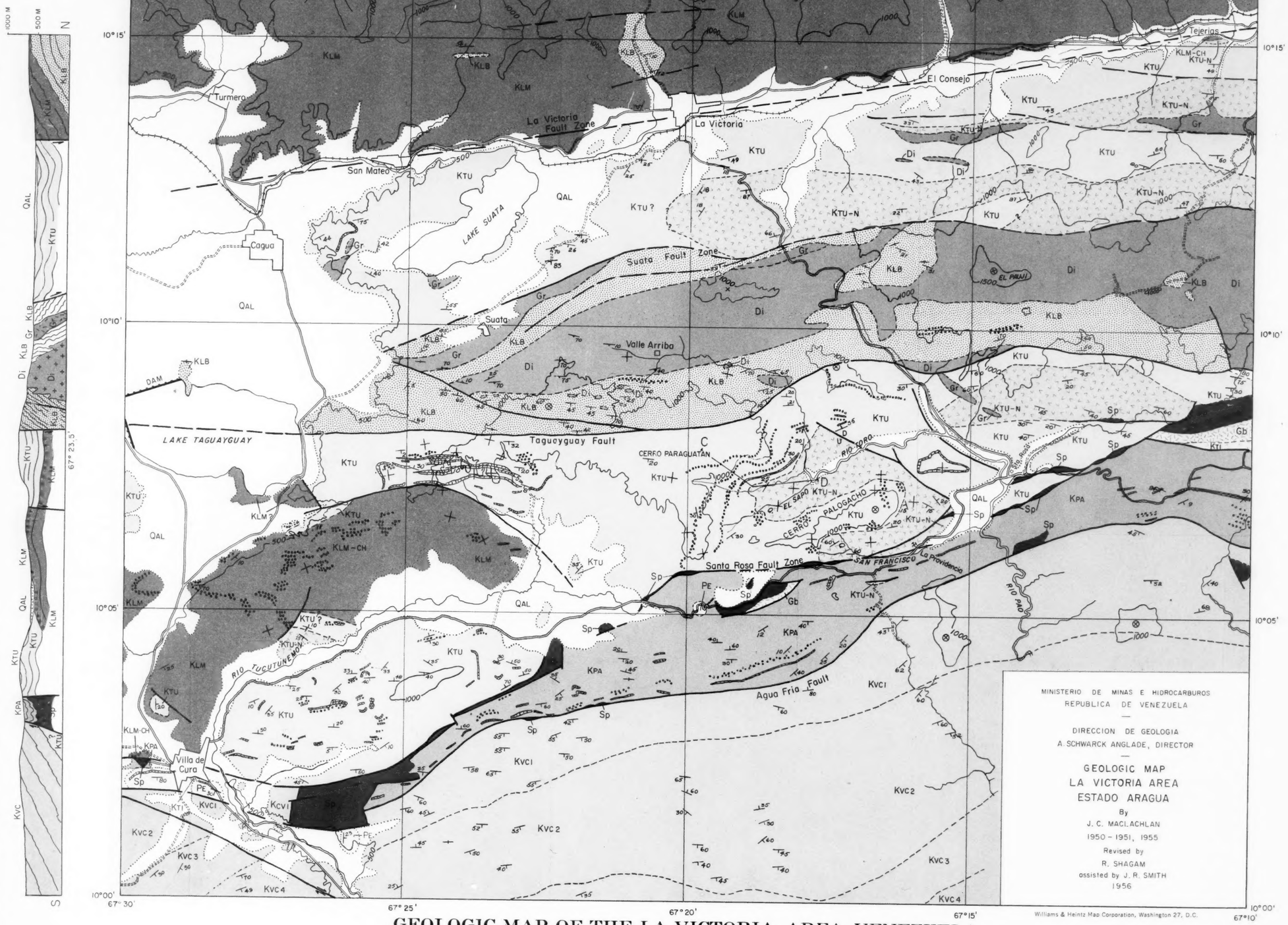
- p — Peridotite with associated pseudogabbro
- gn / a) gnm / b) gni — Hornblende quartz oligoclase gneiss
  - a) metamorphosed to garnet pyroxene hornblende plagioclase gneiss
  - b) intruded by younger granites
- qbs — Quartz-biotite schist

- Strike and dip of bedding
- Strike and dip of foliation
- Strike and dip of axial-plane cleavage
- Mineral lineation, rodding, or intersection of two S-planes
- Minor fold axis
- Thrust fault, barbs on upper plate
- High angle fault





GEOLOGIC MAP OF NORTH-CENTRAL COJEDES, VENEZUELA



LEGEND

- QAL — QUATERNARY ALLUVIUM
- PE — PALEOCENE SEDIMENTARY ROCKS
- KPA — PARACOTOS FORMATION
- KTI — TIARA VOLCANIC ROCKS
- KVC — VILLA DE CURA GROUP
  - KVC4 SANTA ISABEL FORMATION
  - KVC3 EL CARMEN FORMATION
  - KVC2 EL CHINO FORMATION
  - KVC1 EL CANO FORMATION
- KTU — TUCUTUNEMO FORMATION
  - KTU-N — LOS NARANJOS VOLCANIC MEMBER
- KLM — LAS MERCEDES FORMATION
  - KLM-CH — CHARALLAVE CONGLOMERATE MEMBER
- KLB — LAS BRISAS FORMATION

INTRUSIVE IGNEOUS ROCKS

- Di — DIORITE
- Gr — SODA GRANITE
- Gb — GABBRO
- SERPENTINIZED PERIDOTITE

SYMBOLS

- FAULT
- GEOLOGIC CONTACTS
- BOUNDARY OF ALLUVIUM
- CONSPICUOUS CONGLOMERATE BED
- CONSPICUOUS LIMESTONE BED
- STRIKE AND DIP OF BEDDING
- STRIKE AND DIP OF FOLIATION
- PEAK
- 1000 — TOPOGRAPHIC CONTOURS IN METERS

0 1 2 3 4 5 KILOMETERS

MINISTERIO DE MINAS E HIDROCARBUROS
REPUBLICA DE VENEZUELA

DIRECCION DE GEOLOGIA
A. SCHWARCK ANGLADE, DIRECTOR

GEOLOGIC MAP
LA VICTORIA AREA
ESTADO ARAGUA

By
J. C. MACLACHLAN
1950 – 1951, 1955
Revised by
R. SHAGAM
assisted by J. R. SMITH
1956

Williams & Heintz Map Corporation, Washington 27, D.C.

GEOLOGIC MAP OF THE LA VICTORIA AREA, VENEZUELA

67°15'

18°15'

67°07'30"

EXPLANATION

QUATERNARY

Qsd

Surficial deposits

UNCONFORMITY

TERTIARY

Oligo - Miocene — ? Eocene

Tjp

Juana Díaz and Ponce formations undifferentiated

Te

Eocene(?) rocks

UNCONFORMITY

Paleocene

Tj

Jicara formation

CRETACEOUS

Upper Cretaceous — ? Maestrichtian

Ksc / Ksv / Ksa

San Germán formation

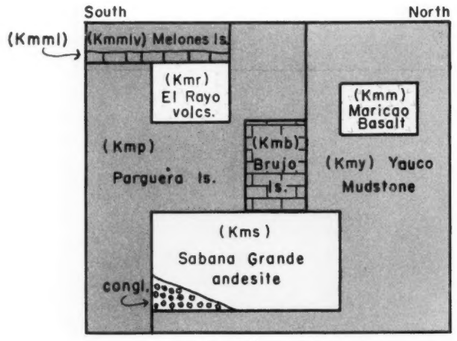
(Ksc) Cotui limestone

(Ksv) andesite

(Ksa) Cabo Rojo agglomerate

UNCONFORMITY

Maestrichtian



Mayagüez Group

? Turonian

Kbl / Kb

Río Blanco formation

(Kbl) Los Marías limestone member

? Cenomanian

Kl

Río Loco formation

Kv

undifferentiated volcanics

?Cretaceous

UNCONFORMITY

Bs — Bc — Ba

Punta Algarrobo

MAYAGÜEZ

Punta Guanajibo

Laguna Joyuda

HORMIGUEROS

Las Mesas - Fraile anticline

Hormigueros syncline

Guanajibo

67°07'30"

67°00'

18°15'

A

A'

B

B'

C

C'

Kmm

Kmy

Kb

Kbl

Kl

Bs

Kms

Kmp

Ksc

Ksv

Osd

Qsd

Ba

2840

2866

2860

2901

2863

2847

2835

2898

2632

2625

2298

2301

2926

2764

2788

2891

2886

2839

2717

2765

LIMÓN

LAS VEGAS

ROSARIO

CONSUMO

Las Vegas syncline

Las Mesas - Fraile anticline

Rosario anticline

Hormigueros syncline

Cordillera fault

U

D

Jicara formation

Ksc, Ksv, Ksa

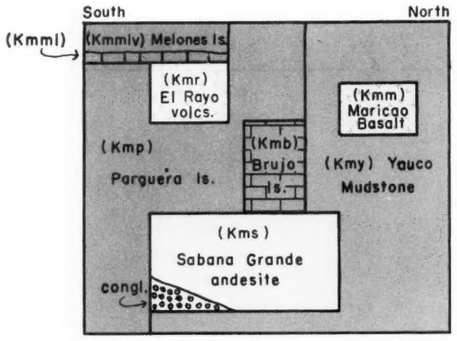
San Germán formation

(Ksc) Cotui limestone

(Ksv) andesite

(Ksa) Cabo Rojo agglomerate

UNCONFORMITY



Mayagüez Group

Kbl, Kb

Río Blanco formation

(Kbl) Los Marías limestone member

Kl

Río Loco formation

Kv

undifferentiated volcanics

?Cretaceous

UNCONFORMITY

Bs — Serpentinite

Bc — Silicified rock

Ba — Spilite, amphibolite

Bermeja complex

Pal

Upper Cretaceous

? Maestrichtian

Maestrichtian

? Turonian

? Cenomanian

CRETACEOUS

18°07'30"

INTRUSIVE IGNEOUS ROCKS

d — diorite porphyry

g — granodiorite

zone of alteration

Fault — inferred fault — covered fault

Thrust fault

Gravity – slide block

Anticline, overturned — Syncline, overturned

Strike and dip of beds, overturned, vertical, horizontal

Punta Guanajibo

Laguna Joyuda

Guanajibo

Las Mesas - Fraile Cabras anticline

Hormigueros syncline

anticline

HORMIGUEROS

VALLE

CABO ROJO

PUERTO REAL

fault

San German

Qsd

Kms

Kmy

Kmp

Kmm

Bs

Ba

Kl

Kv

?Kl

?Kms

?Ksv

Ksc

Ksv

LIMÓN
LAS VEGAS
ROSARIO
HORMIGUEROS
VALLE DE GUANAJIBO
SAN GERMÁN
Las Mesas - Fraile anticline
Cordillera fault
Rosario anticline
Hormigueros syncline
Guanajibo anticline
San Germán fault
18° 07' 30"

UNCONFORMITY

Bs Serpentinite | Bc Silicified rock | Ba Spilite, amphibolite

Bermeja complex

18°07'30"

INTRUSIVE IGNEOUS ROCKS

d diorite porphyry

g granodiorite

zone of alteration

Fault | inferred fault | covered fault

Thrust fault | Gravity – slide block

Anticline, overturned | Syncline, overturned

Strike and dip of beds, overturned, vertical, horizontal

Strike and dip of foliation, vertical, plunge of lineation

C——C Line of structure section

Formation border

2433 Fossil locality | 42 Igneous specimen

Quarry

Laguna Joyuda

Qsd

Kmy

Bs

anticline

Kms

?Kl

?Kms

Kv

CABO ROJO

?Ksv

Ksc

Ksv

E'

fault

German

San

PUERTO REAL

D? U?

Cabo Rojo syncline

?Kmp

Ksa

LLANOS TUNAS

EL BRUJO

Punta Guaniquilla

?Tjp

Tjp

BOQUERÓN

Bahía de Boquerón

18° 07' 30"

VALLE DE GUANAJIBO

VALLE DE LAJAS

Hormigueros syncline

Guanajibo anticline

SAN GERMÁN

LLANOS TUNAS

EL BRUJO

LAJAS

Laguna Cartagena

Tea syncline

La Plata syncline

German fault

Brujo fault

Qsd Kms Kmp Kmy Kmml Kmr Kv Bs Bc Kl Ksv Ksc Ksa Kmb Te Tj

C——————C Line of structure section

Formation border

2433 Fossil locality

42 Igneous specimen

Quarry

Punta Guaniquilla

Punta Melones

Bahía de Boquerón

BOQUERÓN

LLANOS TUNAS

Cabo Rojo syncline

Melones syncline/Kmmlv

Bermeja

Ksv Ksc Ksa Kv ?Kmp Qsd Tjp ?Tjp Bs Bc Ba Kmp Kmmlv

18°00'

LLANOS TUNAS

EL BRUJO

LAJAS

V A L L E D E L A J A S

Laguna Cartagena

Bermeja anticline

Bermeja fault

Parguera fault

Parguera anticline

Vertero syncline

Montalva fault

Tea syncline

La Plata Syncline

PARGUERA

Isla Maguey

Isla Cueva

Isla Guayacán

18° 00'

5°
True North
Magnetic North

18°00'

Punta Melones

Melones syncline

Bermeja anticline

Bermeja fault

CABO ROJO

Bc Bs Ba Kmp Kmmlv Kv Tjp Qsd

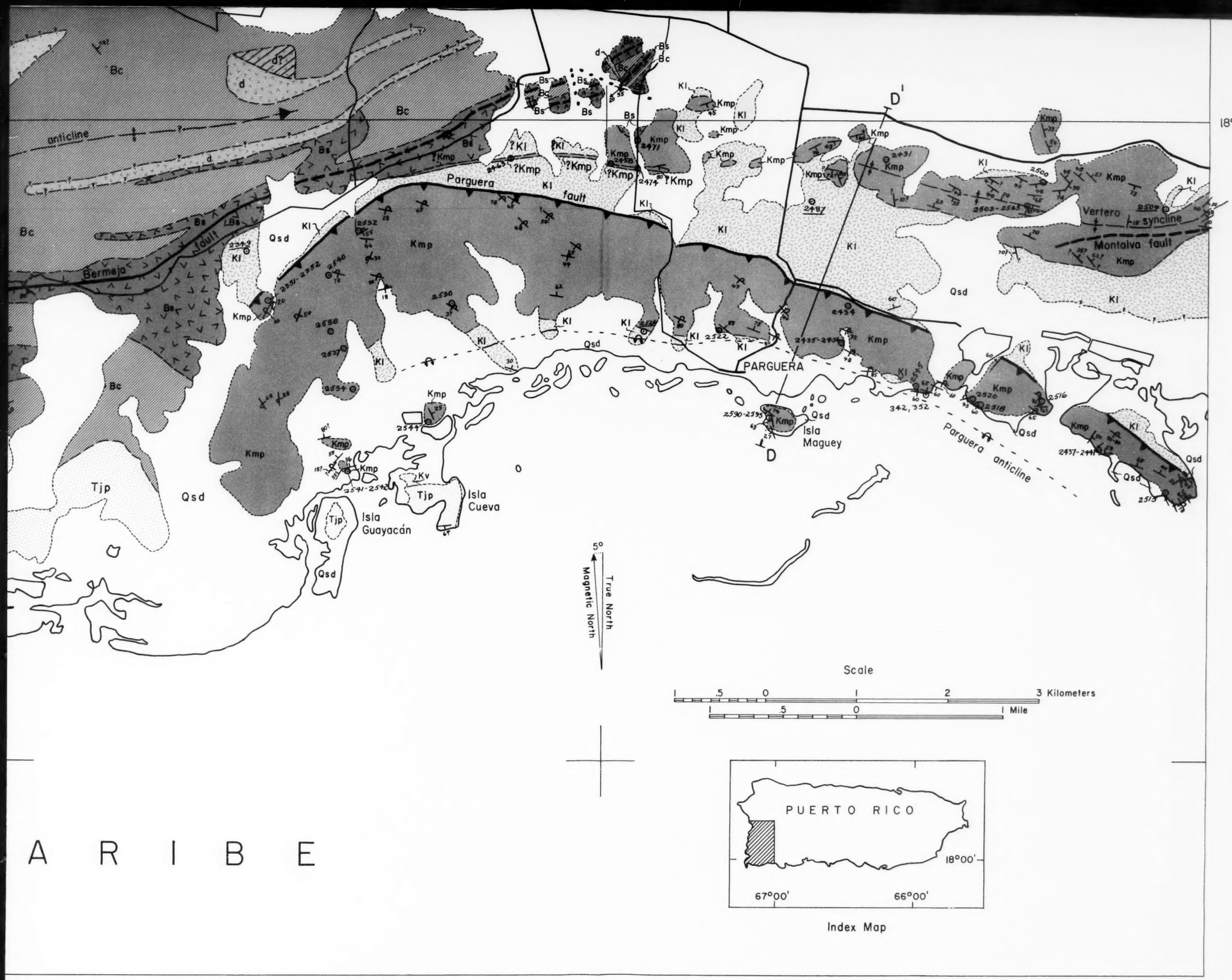
M A R C A R I

67°15'

67°07'30"

GEOLOGIC MAP OF THE MAYAGUEZ AREA

Geology by P. H. Mattson, 1953-1956



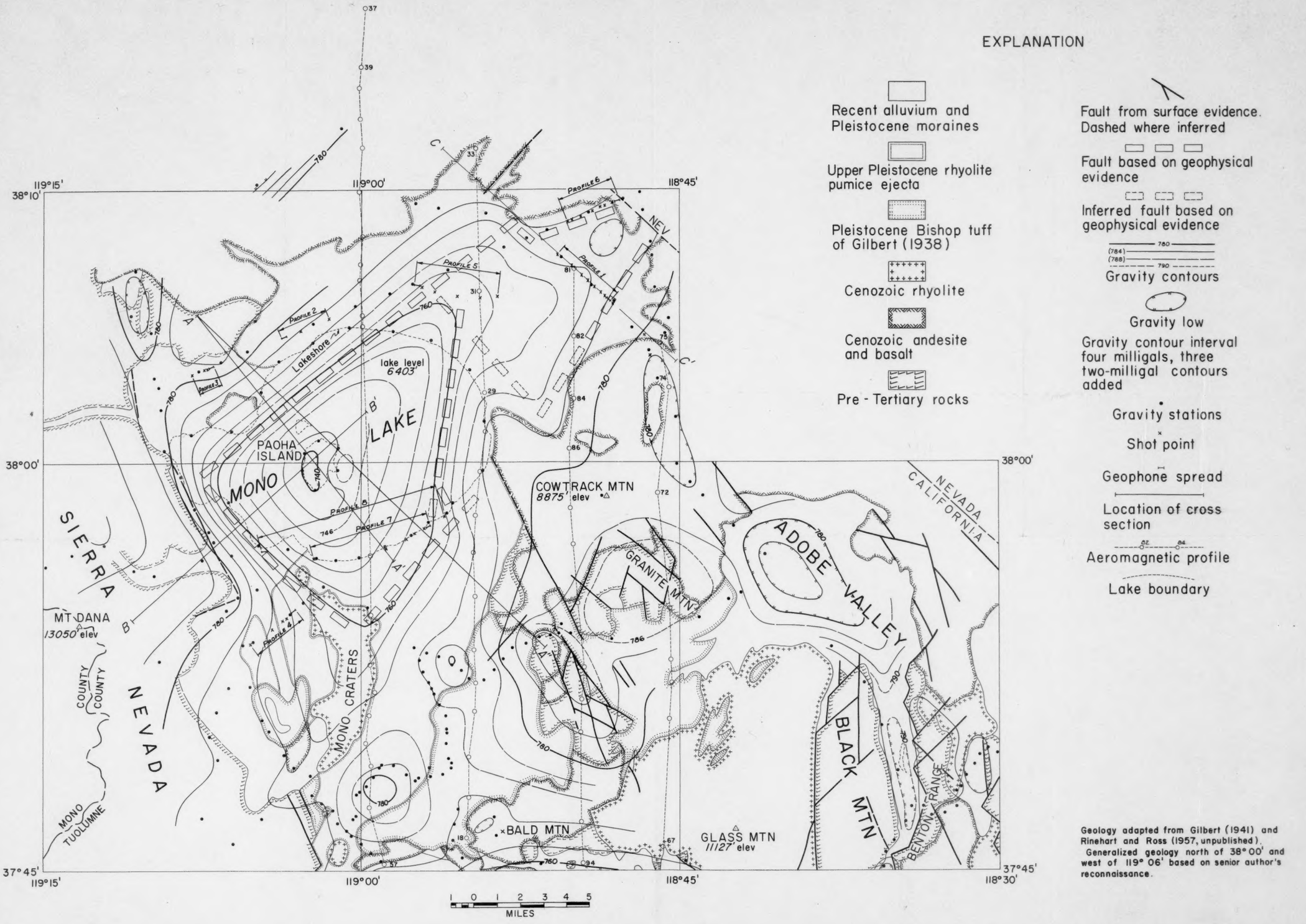
A R I B E

'30"

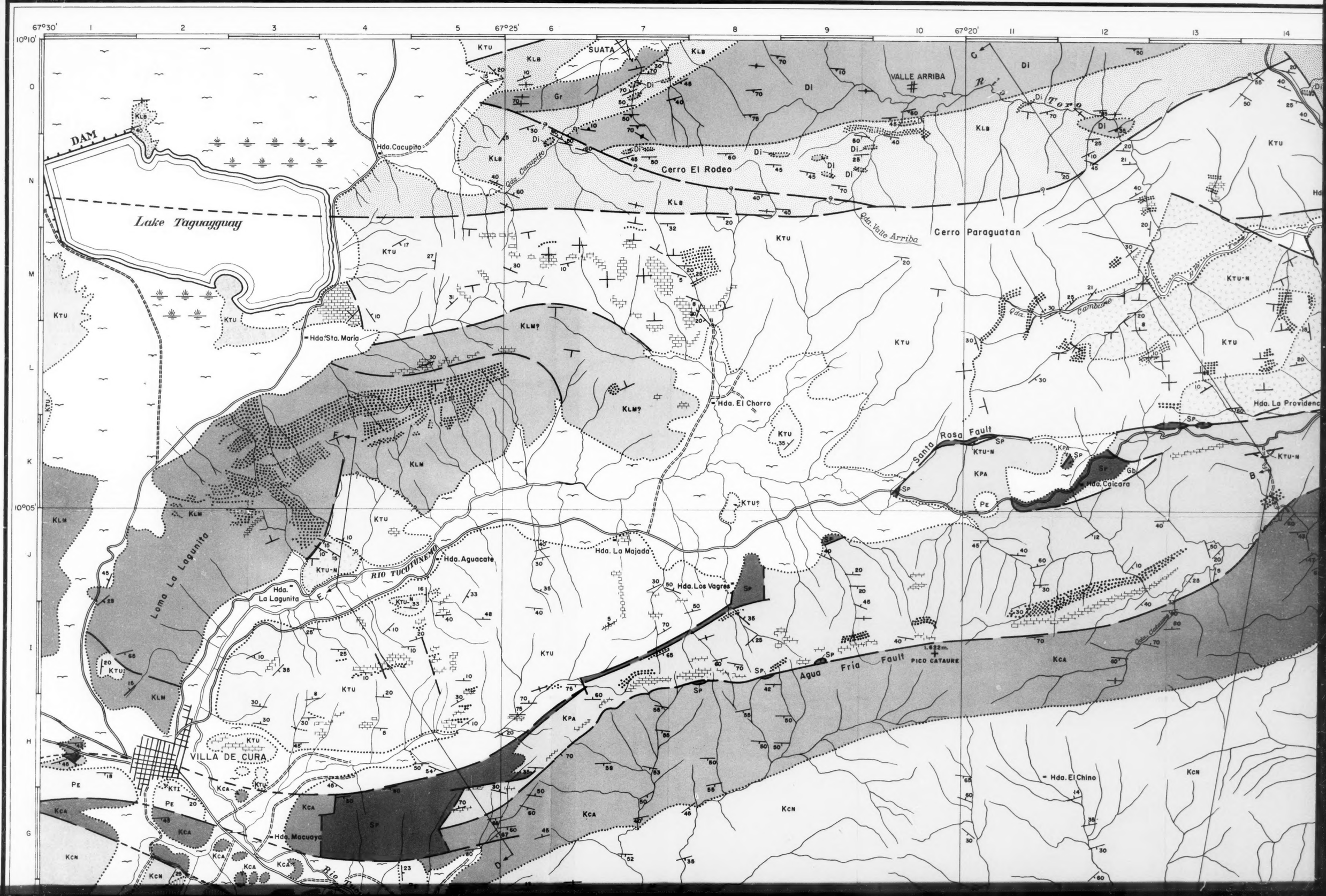
AGUEZ AREA, PUERTO RICO

ttson, 1953-1956

Williams & Heintz Map Corporation, Washington 27, D.C.



COMBINED GRAVITY AND GEOLOGIC MAP OF MONO BASIN, CALIFORNIA, SHOWING LOCATIONS OF SEISMIC AND AEROMAGNETIC PROFILES


67°30'
67°25'
67°20'
10°10'
10°05'
DAM
Lake Taguaygua
Hda. Cacupito
Qda. Cacupito
SUATA
Cerro El Rodeo
VALLE ARRIBA
Qda. Valle Arriba
Cerro Paraguatan
Toro
Qda. Cambure
Hda. La Providenc
Hda. Sta. María
Hda. El Chorro
Santa Rosa Fault
Hda. Caicara
Hda. La Majada
Hda. Aguacate
Hda. La Lagunita
RIO TUCUTUNEMO
Loma La Lagunita
Hda. Los Vagres
Agua Fria Fault
PICO CATAURE
1.622m.
VILLA DE CURA
Hda. Macuayc
Hda. El Chino
Qda. Catauro
KTU
KTU-N
KLB
KLM
KLM?
KPA
KCA
KCN
PE
SP
GR
DI
GB

14 15 67°15' 16 17 18 19 20 67°10' 21 22 23 24 25 67°05'

10°10'

10°05'

0 N M L K J I H G

Hda. El Toro

Hda. La Fundación

Hda. Sta. Lucía

Hda. Sta. Rosa

Hda. La Providencia

Hda. El Paito

Hda. Agua Fría

Hda. El Caño

Río Santa Lucía

Río Pao

Río Camaguán

Qda. Los Jardines

Qda. Cumbote

Qda. Negrito

Agua Fría

Loma de Hierro

TIARA

Fault

Di

KLB

KTU

KTU-N

KPA

KCA

KCN

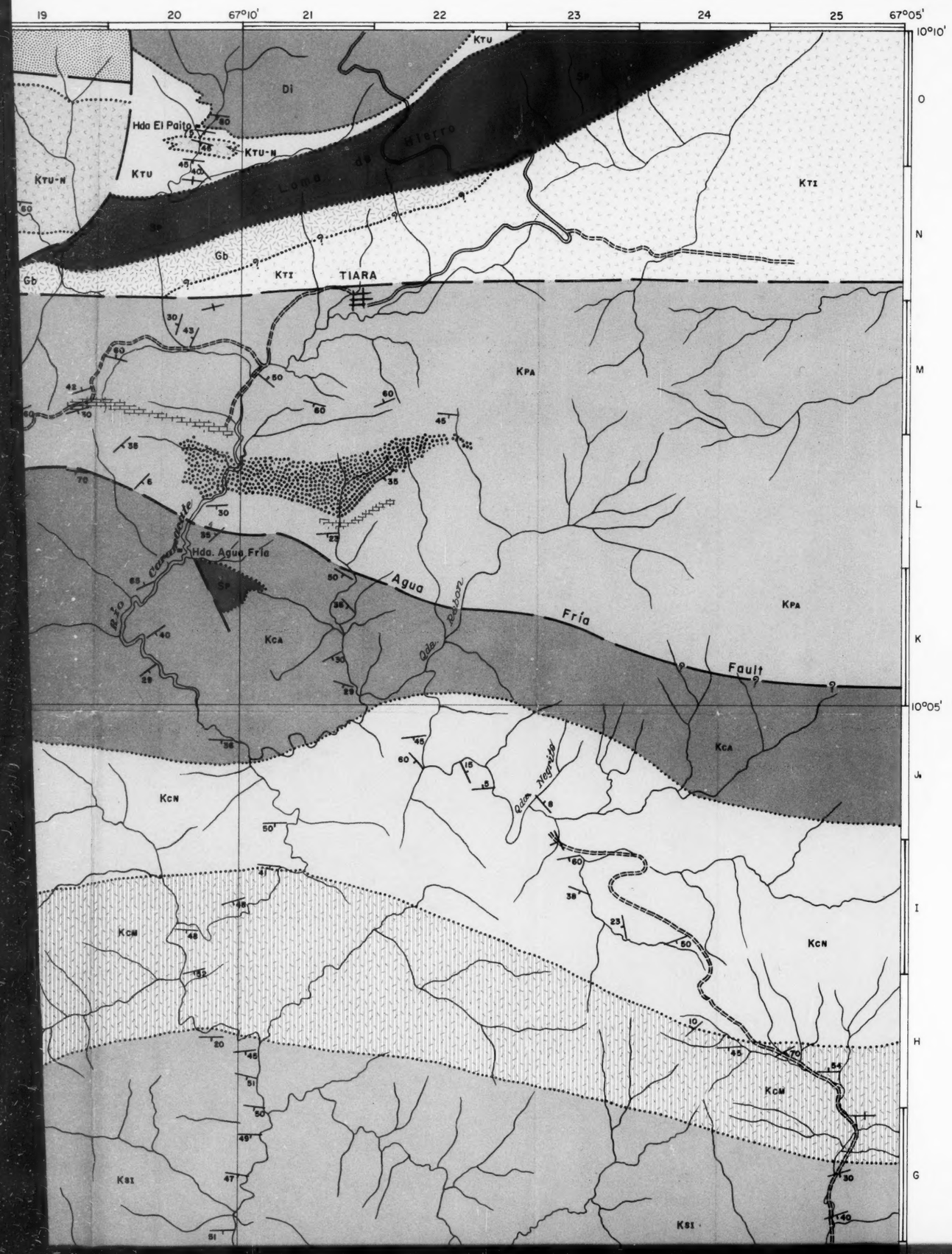
KCM

KSI

KTI

Gb

SP



GEOLOGIC MAP OF CENTRAL ARAGUA, VENEZUELA.

SCALE 1:50,000

1 0 1 2 3

KILOMETERS

LEGEND

MOUNTAIN BELT

ALLUVIUM

PALEOCENE— *Brown shales, siltstones, gray limestones.* — Pe

PARACOTOS formation—Maestrichtian— *Gray aphanitic limestone, conglomerate of basalt pebbles, carbonaceous phyllitic shales.* — KPA

TIARA VOLCANIC ROCKS—*Spilitic basalts and volcanic sedimentary rocks.* — KTI / Gb.

VILLA DE CURA GROUP (Age? Pre-Turonian)

- SANTA ISABEL fm. — KSI — *Quartz-albite granulites, epidote-chlorite-quartz-albite schists.*
- EL CARMEN fm. — KCM — *Porphyritic spilitic metabasalts and volcanic metasedimentary rocks.*
- EL CHINO fm. — KCN — *Aphanitic chlorite-quartz-albite rocks, black laminated cherts, spilitic metabasalts, etc.*
- EL CANO fm. — KCA — *Aphanitic laminated metatuffs and minor basic to andesitic volcanic rocks.*

CARACAS GROUP (Age? possibly Early Cretaceous)

- TUCUTUNEMO fm. — KTU / KTU-N — *Black fine-grained limestone, feldspathic quartz conglomerate, carbonaceous phyllite. Metabasalts and aphanitic laminated metatuffs— Los Naranjos member.*
- LAS MERCEDES fm. — KLM — *Quartz conglomerate— Charallave member. Calcareous graphitic schists and phyllites, minor fine-grained graphitic limestones.*
- LAS BRISAS fm. — KLB — *Quartz-muscovite schist. Scattered granitic, quartzite, vein quartz pebbles.*

SODA GRANITE — Gr

DIORITE — Di

SERPENTINIZED PERIDOTITE PYROXENITE — PK

GUARICO VALLEY

ALLUVIUM

PALEOCENE—MAESTRICHTIAN— *Morro (reef) limestone.* — Pe — Paleocene

Brown calcareous lithic wacke siltstones, sandstones, and conglomerate; bedded chert; blue-green aphanitic limestone. — KM — Maestrichtian

ARRAYANES formation—Coniacian— *Blue lithic wacke siltstones, sandstones, and conglomerates.*

EL TOTUMO—Turonian— *Black aphanitic foraminiferal limestone, and brown shales.* — KTO

TIARA VOLCANIC ROCKS — KTI / Gb.

SERPENTINIZED PERIDOTITE

Formation Boundaries

Trace and dip of faults (up and down) — U / D 45

Postulated faults under alluvium

TOWNS

MAIN ROADS

JEEP ROADS

DRAINAGE

Strike and dip of bedding — 60

Strike and dip of foliation — 40

Vertical bedding and foliation

VILLA DE CURA
Hda. Macuaya
Río Tucutunemo
Guárico Fault
Hda. Tierra Blanca
Hda. El Carmen
Hda. Caravanelo
Hda. Sta. Isabel
Hda. El Chacao
Qda. El Semen
Qda. Piritu
Qda. La Puerta
Qda. Las Minas
Qda. San Juan Morado
RIO GUARICO
KTO
KTI
KTU
KPA
KCA
KCN
KCM
KSI
KM
PE
PX
9°55'
67°30'
67°25'
67°20'
SOUTH
METERS 1.200
800
400
0
A
B
SP
GEOLOG

GEOLOGIC MAP AND CROSS SECTION OF CENTRAL ARAG

TOWNS

MAIN ROADS

JEEP ROADS

DRAINAGE

(up and down)

Postulated faults under alluvium

Strike and dip of bedding

Strike and dip of foliation

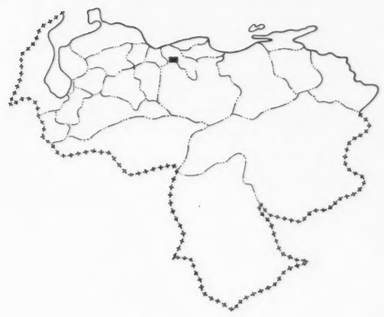
Vertical bedding and foliation

Strike and dip of bedding where dip is > 5°

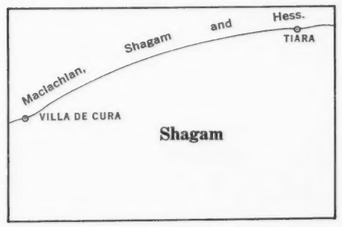
Horizontal bedding

Cross section lines

GEOGRAPHIC LOCATION



GEOLOGY BY:



FIELD WORK SPONSORED BY: DIRECCION DE GEOLOGIA, MINISTERIO DE MINAS E HIDROCARBUROS, REPUBLICA DE VENEZUELA

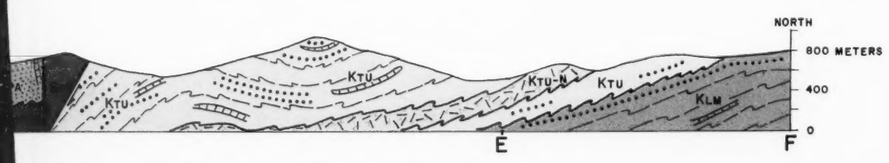
R. SHAGAM

GEOLOGICAL CARTOGRAPHY — DRAFTING SECTION

DRAFTED BY: R. A. RAMIREZ R.—

1957

Williams & Heintz Map Corporation, Washington 27, D.C.

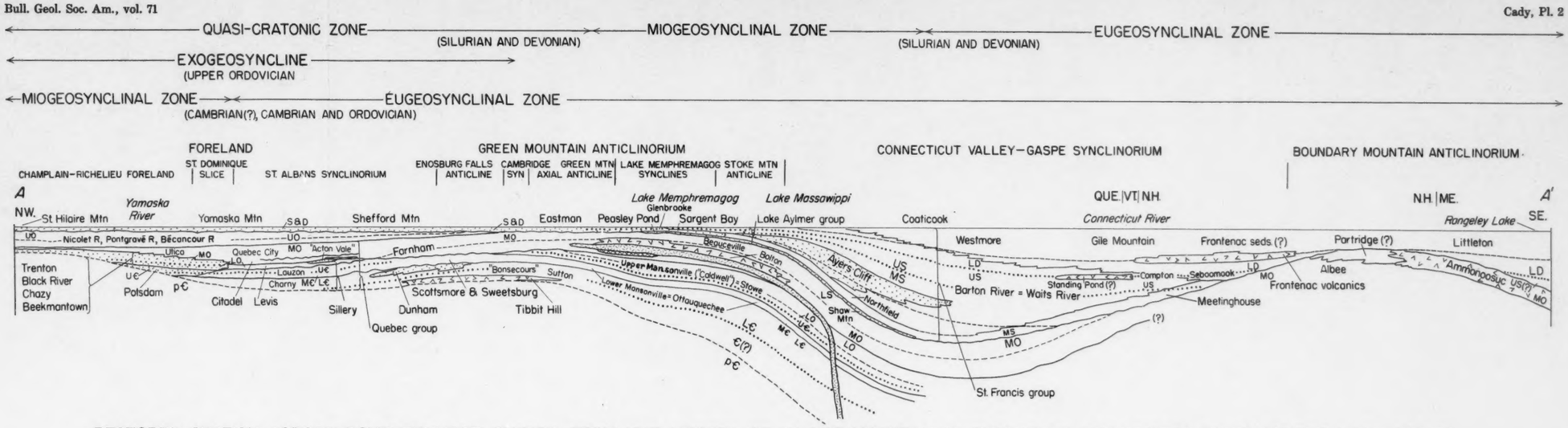


SECTIONS ABC, DEF, GH.

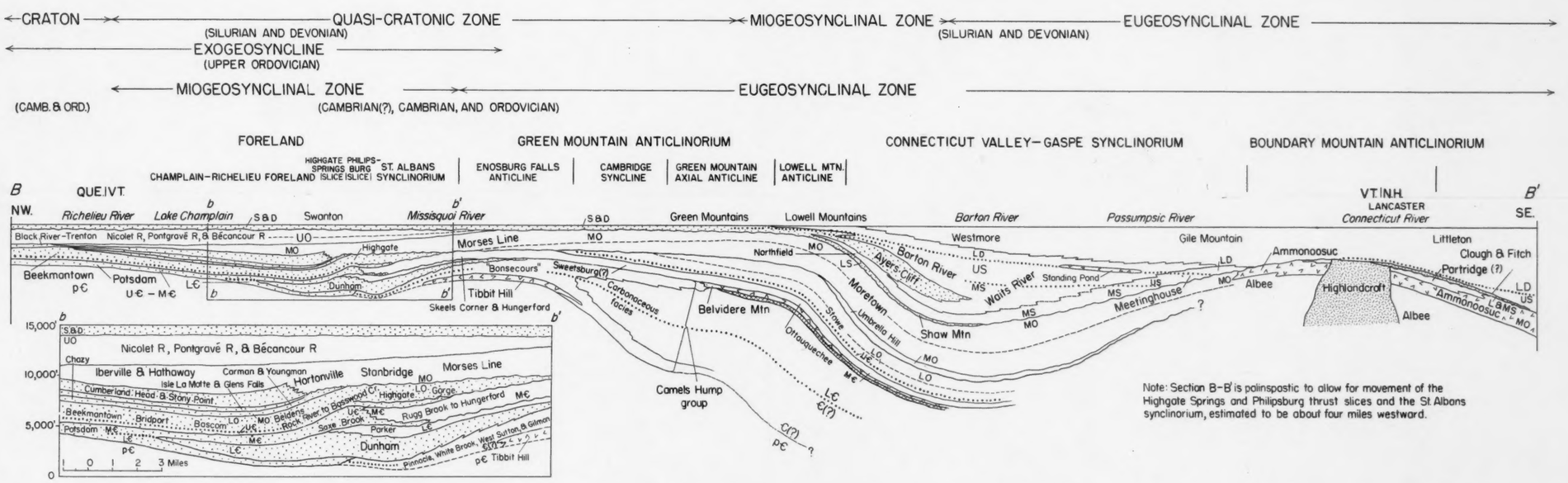
Vertical scale = Horizontal scale = Map scale

Legend as for map

REA, VENEZUELA



RESTORED SECTION ACROSS SOUTHERNMOST QUEBEC, FROM THE VICINITY OF HILAIRE MOUNTAIN IN QUEBEC TO THE VICINITY OF RANGELEY LAKE, MAINE



RESTORED SECTION ACROSS NORTHERN VERMONT, BETWEEN THE HEAD OF THE RICHELIEU RIVER, QUEBEC AND VICINITY OF LANCASTER, NEW HAMPSHIRE

Note: Section B-B' is palinspastic to allow for movement of the Highgate Springs and Philipsburg thrust slices and the St Albans synclinorium, estimated to be about four miles westward.

EXPLANATION

Carbonate-quartzite assemblage*

Graywacke-shale assemblage*

Volcanic rocks

Mafic to felsic intrusive rocks

Ultramafic intrusive rocks

Time surfaces — Located at series boundaries

Form lines — Time surfaces centrally located in stratigraphic unit to show convergence and relationship to surfaces of unconformable overlap.

*Contain units of differing assemblage that are too thin to show or are of indistinct lithology.

D Devonian
S Silurian
O Ordovician
C Cambrian
p€ Precambrian
L, M, U, Lower, Middle, Upper (Applies to any Paleozoic Unit)

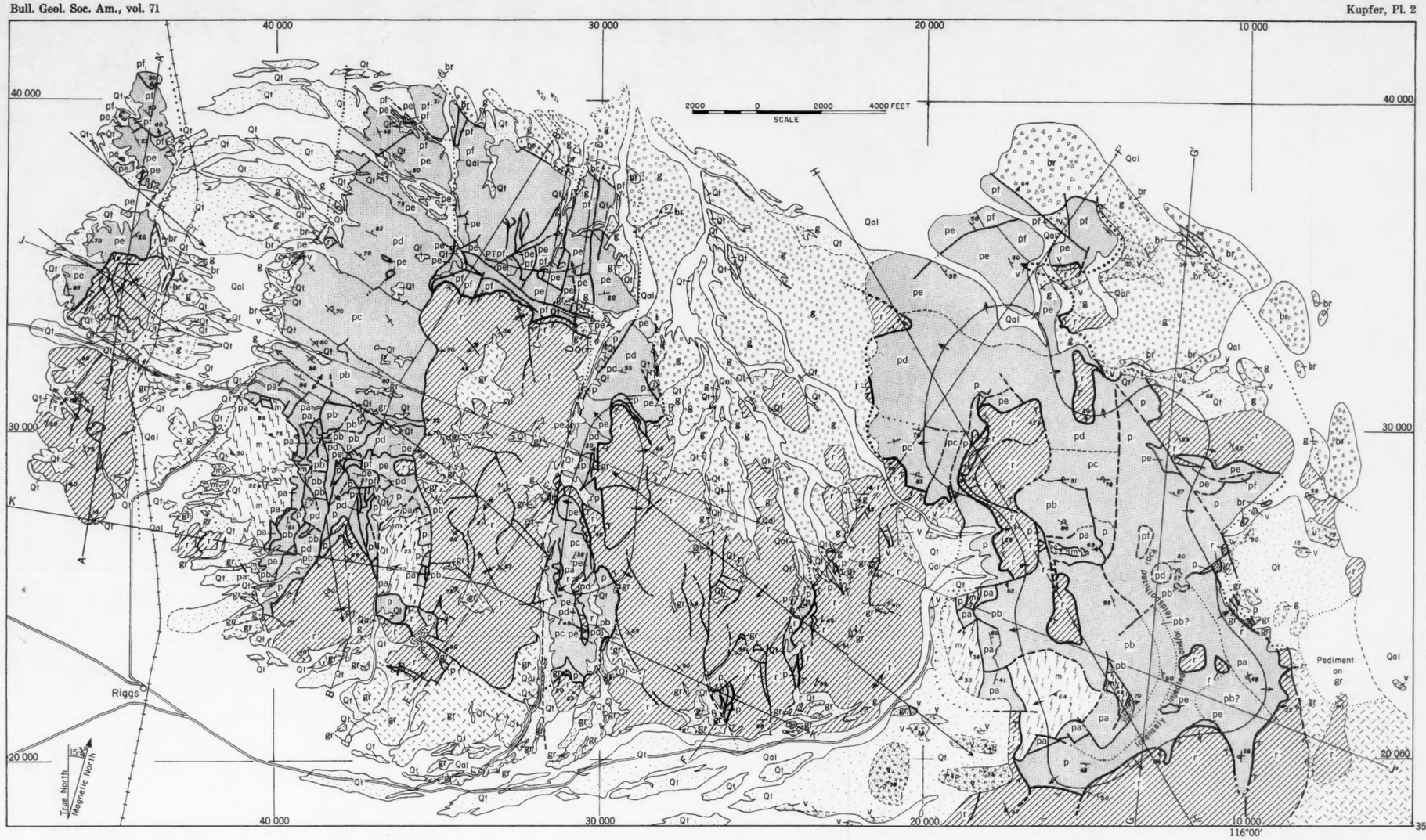
5 0 5 10 15 20 Miles
Approximate horizontal scale

20,000 feet
10,000
0
Vertical scale
Exaggered about twice

W.M. CADY, 1959

RESTORATION OF THE BEDDED ROCKS AND INTRUSIVE IGNEOUS ROCKS OF NORTHERN VERMONT AND NEW HAMPSHIRE AND NEARBY PARTS OF QUEBEC AND MAINE, AT THE END OF THE EARLY DEVONIAN AND BEFORE THE ACADIAN OROGENY

See Plate 1 for location of restored sections. Sections are transverse to the trend of the orthogeosynclinal belt. Vertical scale is exaggerated about twice.



EXPLANATION

CENOZOIC
- Qal Alluvium
- Qt Terrace gravels
- UNCONFORMITY
- g Older gravels (Tertiary)
- UNCONFORMITY
- br Megabreccia
- UNCONFORMITY
- v Volcanic and sedimentary rocks

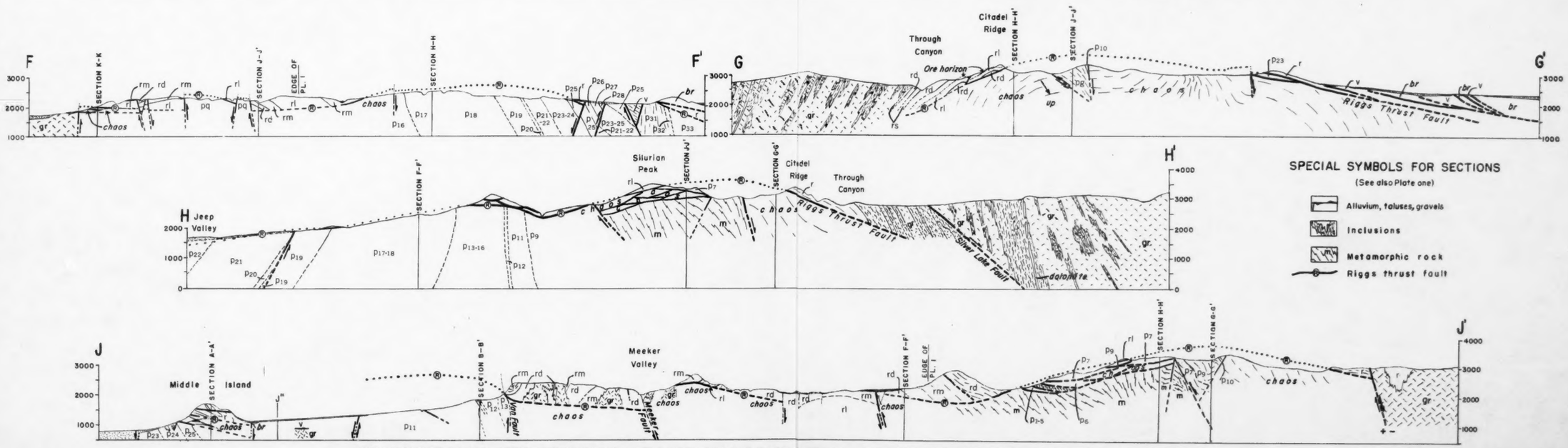
PALEOZOIC (?)
- UNCONFORMITY
- Riggs formation, mainly recrystallized carbonate rock

PRECAMBRIAN PAHRUMP GROUP
- UNCONFORMITY
- p Undifferentiated units
- pf Quartzite and siltstone (26–35)
- pe Mixed sediments (19–25)
- pd Conglomerate and fanglomerate (17–18)
- pc Well-bedded sediments (13–16)
- pb Quartzite with carbonate rocks (7–12)
- pa Mixed sediments (1–6)
- UNCONFORMITY
- m Metamorphic rocks

- gr Granitic rock and inclusions
- Geologic contact
- Fault, showing dip
- Riggs thrust fault

(Dashed where approximate, dotted where hypothetical)

- Dip and strike of inclined and vertical bedding
- Overturned bedding
- Dip and strike of foliation
- Synclinal axis showing plunge. PT=fold does not affect thrust plate.
- Anticlinal Axis
- Dome in thrust plate



SPECIAL SYMBOLS FOR SECTIONS
(See also Plate one)

- Alluvium, taluses, gravels
- Inclusions
- Metamorphic rock
- Riggs thrust fault

GEOLOGIC MAP AND SECTIONS OF THE SILURIAN HILLS, SAN BERNARDINO COUNTY, CALIFORNIA

Modification of Map Sheet 19 (Kupfer, 1954) by permission of California Division of Mines. *See* Plate 1 for cross sections A-A', B-B', C-C', D-D', E-E', and K-K'.

LEGEND

MAESTRICHTIAN – PALEOCENE

4b Massive Morro Limestone and Bedded Flank Limestone } Paleocene

4 Calcareous Siltstone, Sandstone and Conglomerate, Bedded Chert

4a Aphanitic Blue-green Limestone } Maestrichtian

4 Calcareous Siltstones and Sandstones, and Bedded Chert

Unconformity

TIARA VOLCANIC ROCKS (AGE ?)

3 Porphyritic Basalt

Unconformity

2 Metadiorite intrusions

SANTA ISABEL FORMATION (AGE ?)

1 Quartz-albite Granulites, Chlorite Schists, etc. Glaucophane - Epidote - Chlorite - Quartz - Albite Schists

1c Laminated Black Chert

1b Volcanic Conglomerate

1a Basic Metatuffs, Porphyritic Spilitic Metabasalt, Basic Metatuffs

1 Quartz-albite Granulites, Chlorite Schists, etc. Glaucophane-Epidote-Chlorite-Quartz-Albite-Schists

Strike and Dip of Foliation Planes

Faults

Contours at 100m. intervals

Surveyed points and Altitudes

Strike and dip of thrust fault at E-8

0 100 200 300 400 500 600 800 1000 METERS

Field work sponsored by Direccion de Geologia, Ministerio de Minas e Hidrocarburos, Republica de Venezuela

R. SHAGAM

Mts de Cura se Kms.

Qda. Las Minas

HDA. STA. ISABEL

Río Guárico

LA PUERTA

Río Guárico

Qda. Guaiquera

MOSTLY ALLUVIUM COVERED

MOSTLY ALLUVIUM COVERED

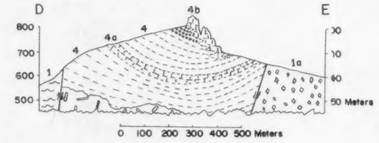
SAN SEBASTIAN 23 Km

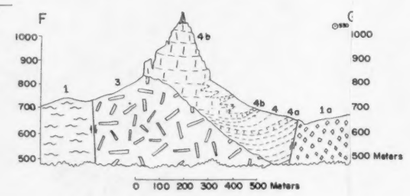
Statue of San Juan

SAN JUAN DE LOS MORROS

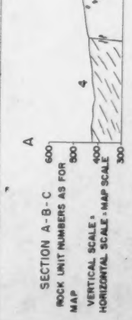
Northeast guard-tower of Penitentiary

SECTIONS D-E and F-G
ROCK UNIT NUMBERS AS FOR MAP
VERTICAL SCALE = HORIZONTAL SCALE = MAP SCALE





SECTION A-B-C
ROCK UNIT NUMBERS AS FOR MAP
VERTICAL SCALE = HORIZONTAL SCALE = MAP SCALE



A - MORROS OF SAN JUAN AND ENVIRONS, GUARICO.

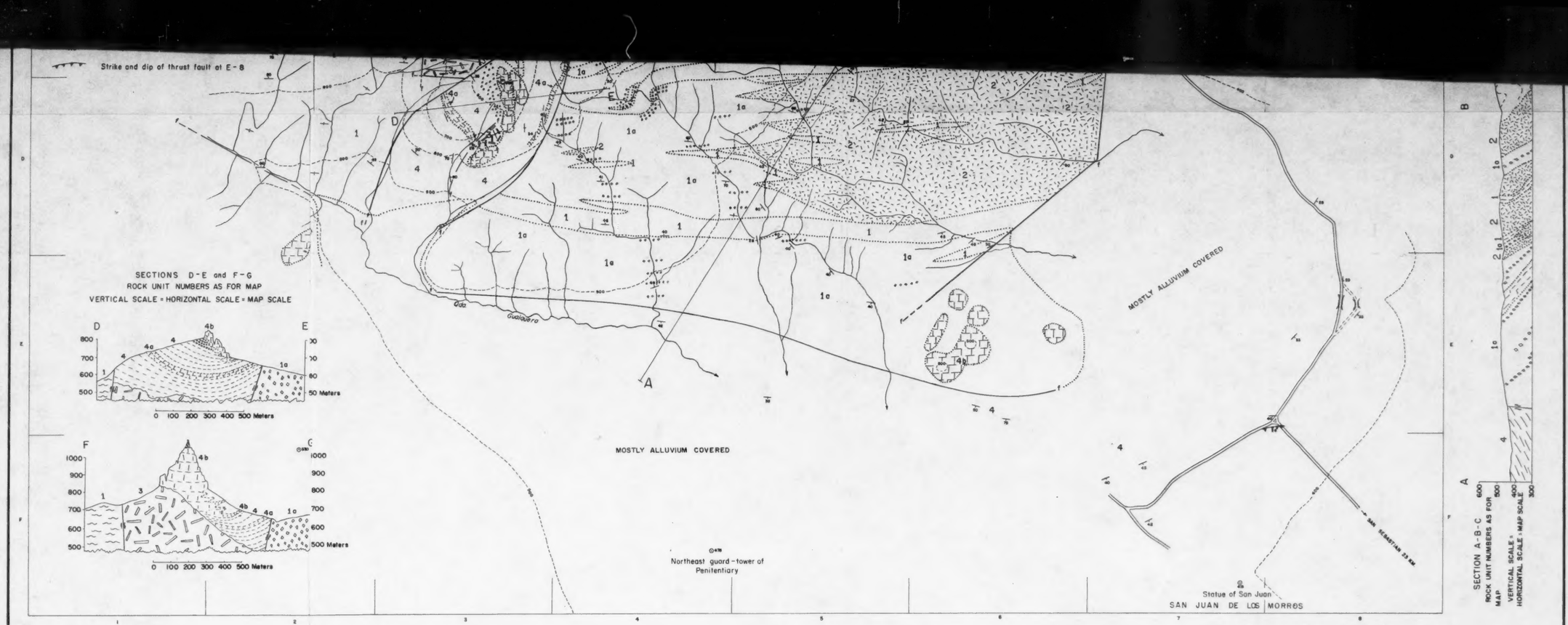
0 100 300 500 700 1000 METERS

LEGEND

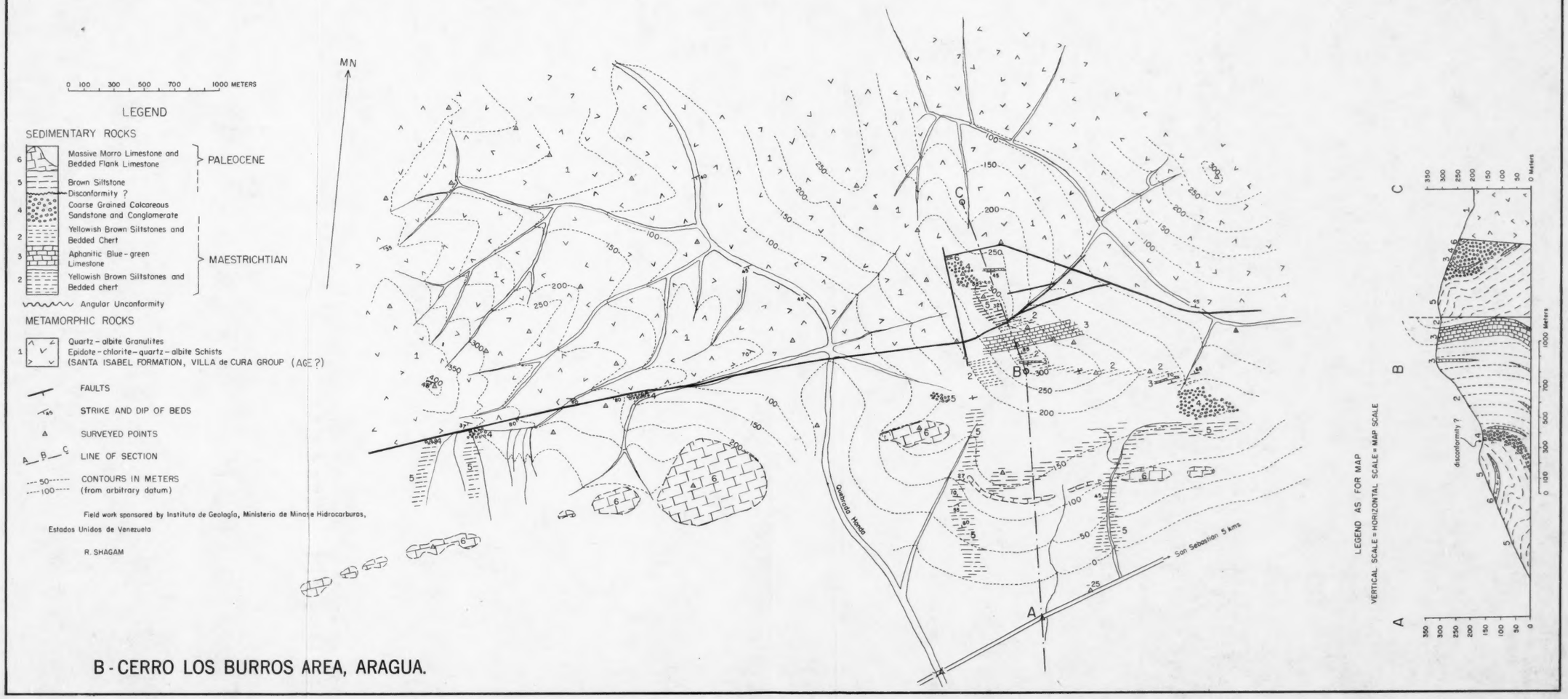
SEDIMENTARY ROCKS

6 Massive Morro Limestone and Bedded Flank Limestone } PALEOCENE

MN

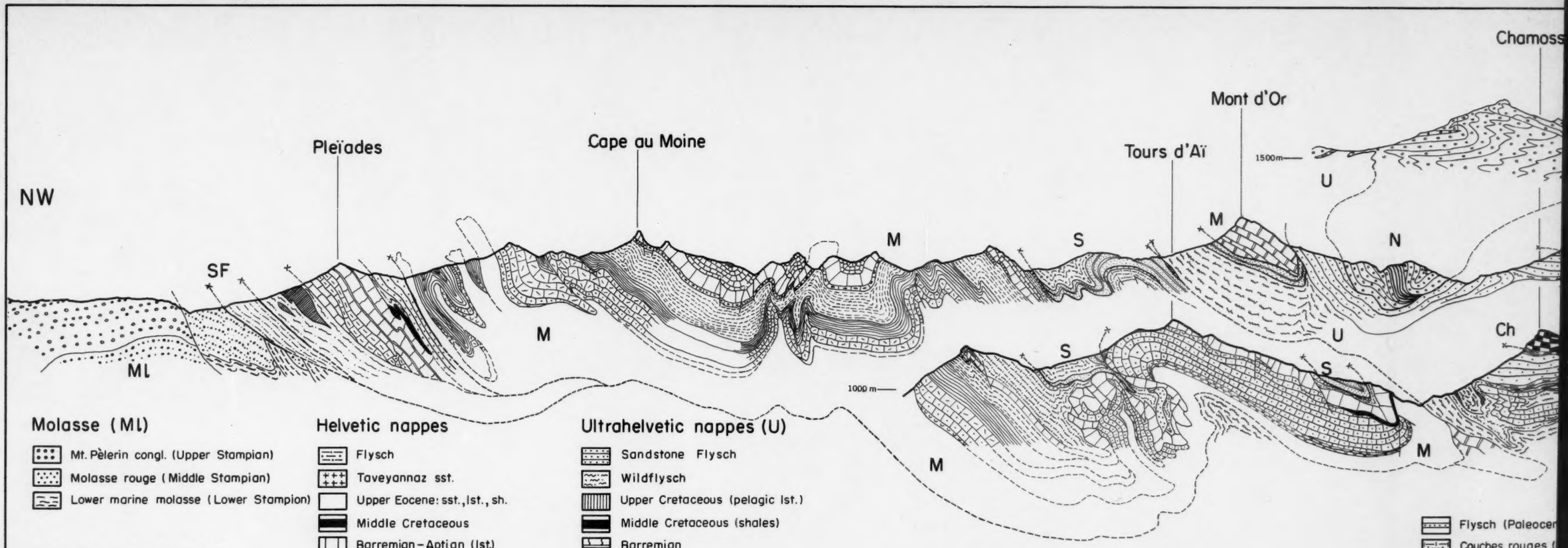

Strike and dip of thrust fault at E-8
SECTIONS D-E and F-G
ROCK UNIT NUMBERS AS FOR MAP
VERTICAL SCALE = HORIZONTAL SCALE = MAP SCALE
D
E
F
G
0 100 200 300 400 500 Meters
50 Meters
MOSTLY ALLUVIUM COVERED
Northeast guard-tower of Penitentiary
Statue of San Juan
SAN JUAN DE LOS MORROS
SECTION A-B-C
ROCK UNIT NUMBERS AS FOR MAP
VERTICAL SCALE = HORIZONTAL SCALE = MAP SCALE

A - MORROS OF SAN JUAN AND ENVIRONS, GUARICO.



B - CERRO LOS BURROS AREA, ARAGUA.

GEOLOGIC MAPS OF MORROS OF SAN JUAN AND ENVIRONS AND CERRO LOS BURROS AREA, VENEZUELA



Molasse (Ml)

- Mt. Pèlerin congl. (Upper Stampian)
- Molasse rouge (Middle Stampian)
- Lower marine molasse (Lower Stampion)

Subalpine Flysch (SF)

- Paleocene? or Lower Oligocene ?

Wi Wildhorn nappe } Helvetic
Di Diablerets nappe }
Mo Morcles nappe, Subhelvetic

Helvetic nappes

- Flysch
- Taveyannaz sst.
- Upper Eocene: sst., lst., sh.
- Middle Cretaceous
- Barremian–Aptian (lst.)
- Lower Barremian shales
- Hauterivian (cherty lst.)
- Up. Valanginian limestones
- Low. Valanginian (sh. & lst.)
- Upper Jurassic limestones
- Middle Jurassic (shales etc.)

Ultrahelvetic nappes (U)

- Sandstone Flysch
- Wildflysch
- Upper Cretaceous (pelagic lst.)
- Middle Cretaceous (shales)
- Barremian
- Hauter.– Valanginian
- Upper Jurassic limestones
- Bajocian to Oxfordian, shales
- Aalenian (black shales)
- Lias (lst. & shale)
- Upper Trias: evaporites, shale, dolomite.

Chamossaire nappe (Ch)

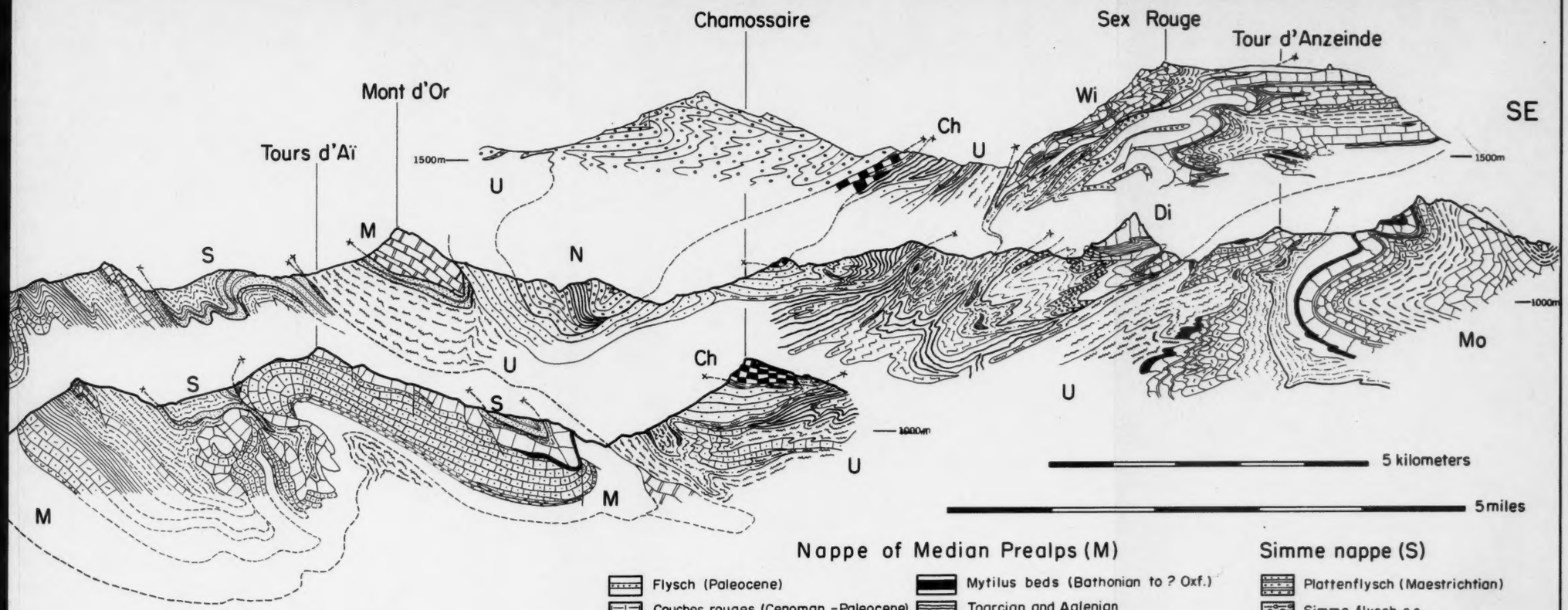
- Aalenian (black shales)
- Lias (limestones)
- Upper Trias (cornieules)

Niesen nappe (N)

- Upper Cretaceous flysch
- Middle Jurassic (shales)
- Upper Trias (cornieules)

- Flysch (Paleocen
- Couches rouges (
- Lower Cretaceou
- Upper Jurassic li
- Argovian nodula
- Zoophycos Dog (Bajocian to Ox

CROSS SECTIONS THROUGH THE PREALPS EAST OF THE RHONE VALLEY



Nappe of Median Prealps (M)

- Flysch (Paleocene)
- Couches rouges (Cenoman.–Paleocene)
- Lower Cretaceous (lst.& chert)
- Upper Jurassic limestones
- Argovian nodular lst.
- Zoophycos Dogger (Bajocian to Oxfordian)
- Mytilus beds (Bathonian to ? Oxf.)
- Toarcian and Aalenian
- Middle Lias
- Lower Lias
- Rhaetic
- Upper Trias (dolom., anhydr., sh.)
- Middle Trias (lst. & dol.)

Simme nappe (S)

- Plattenflysch (Maestrichtian)
- Simme flysch s.s. (Cenomanian–Senonian)

× indicates thrust

mossaire nappe (Ch)

- Aalenian (black shales)
- Lias (limestones)
- Upper Trias (cornieules)

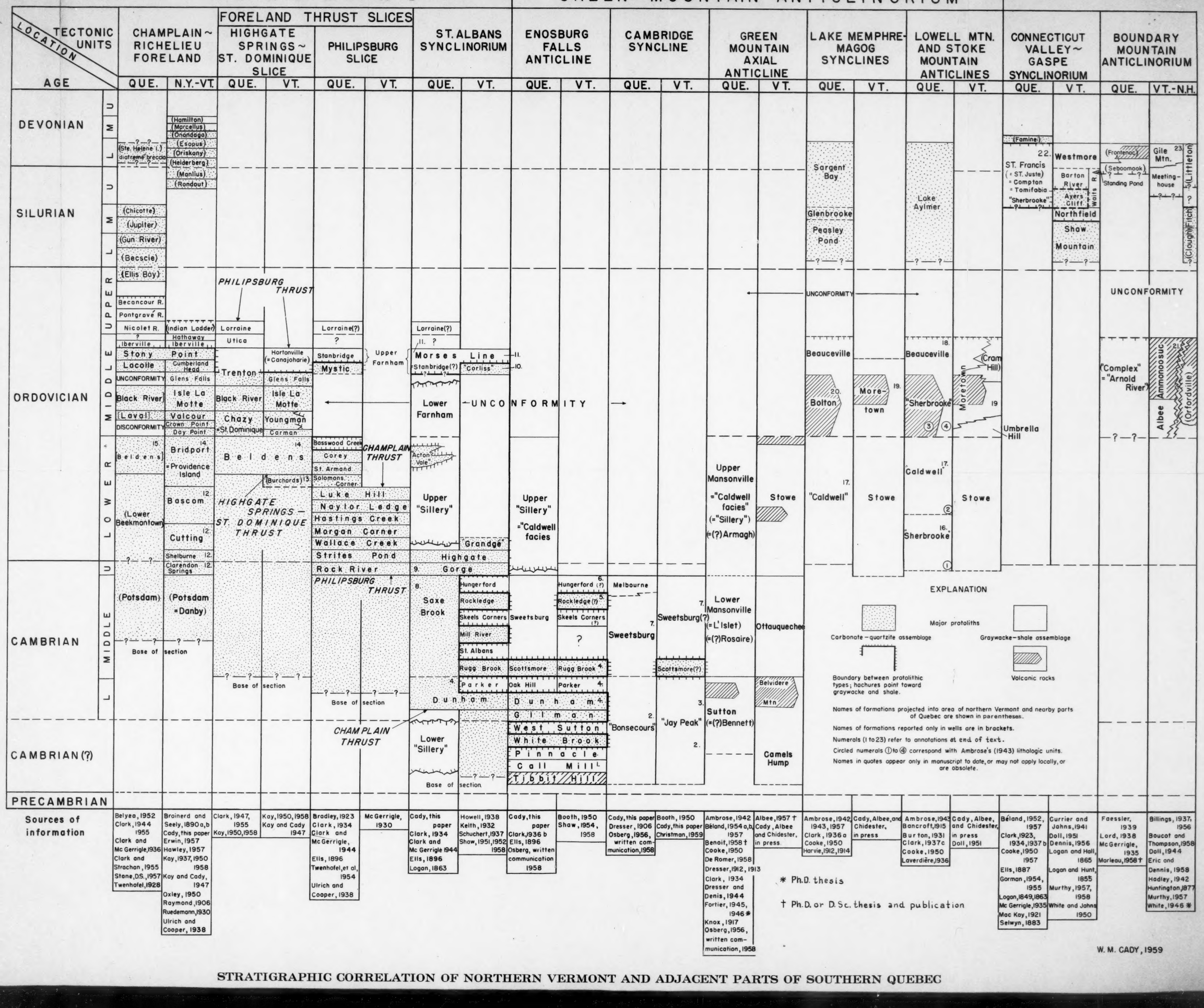
Niesen nappe (N)

- Upper Cretaceous flysch
- Middle Jurassic (shales)
- Upper Trias (cornieules)

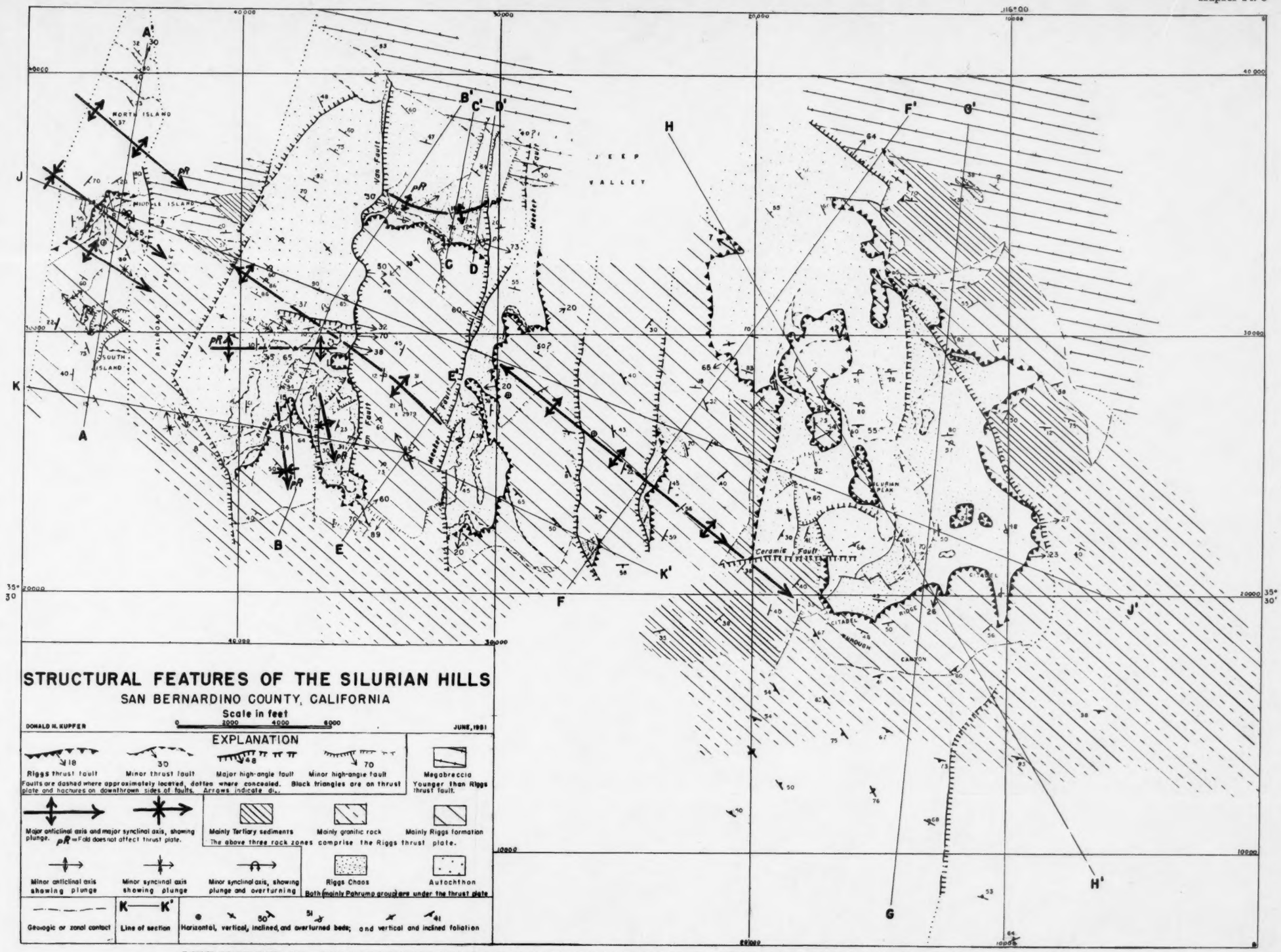
drawn by Walter Martin †

NS THROUGH THE PREALPS EAST OF THE RHONE VALLEY

STRATIGRAPHIC CORRELATION OF NORTHERN VERMONT AND ADJACENT PARTS OF SOUTHERN QUEBEC



| Sources of information | | |



STRUCTURAL FEATURES OF THE SILURIAN HILLS, SAN BERNARDINO COUNTY, CALIFORNIA

Río Tirgua
Manrique Thrust Fault
Cerro La Alegría
Casupa
Vallecito
Cerro Tetas
Flanquillo
Río de Cardozo
Qal
ph
pg
d
ds
gn

Las Mercedes

Casupo

El Tigre

Morrique Thrust Fault

Qal

ph

ds

pg

ad

gn

d

SECTION A–A'

LEGEND

	Qal	Alluvium
Cretaceous	ph	Phyllite with limestone lenses
	ad	Areas of abundant "acid dikes"
	gn	Garnet-pyroxene-amphibole-plagioclase gneiss (meta hornblende-quartz-oligoclase gneiss)
Cretaceous?	pg	Psuedogabbro
	Py	Pyroxenite

N

30°

GEOLOGIC MAP OF THE TINAQU

NAQUILLO PERIDOTITE, VENEZUELA

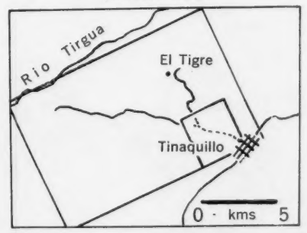
LEGEND

- Qal — Alluvium
- Cretaceous: ph — Phyllite with limestone lenses
- Areas of abundant "acid dikes"
- gn — Garnet-pyroxene-amphibole-plagioclase gneiss (meta hornblende-quartz-oligoclase gneiss)
- Cretaceous?: pg — Psuedogabbro
- Py — Pyroxenite
- d — Dunite, showing serpentinized zone
- Contact
- Fault
- Major thrust fault
- Amphibole- or enstatite-rich layers
- Schistosity
- Joints
- Large oriented orthopyroxenes
- Lineation imparted by mineral parallelism, axes of small folds, or rarely the intersection of S-planes.
- Cross fiber bearing planes.
- Foliation (in section)

N

3°

INDEX MAP



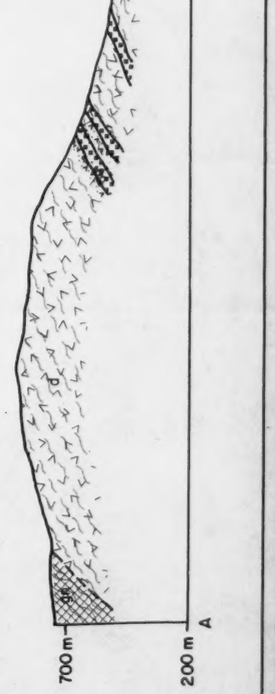
Base Map: 1:10,000 Plane table sheet extended to northeast by uncontrolled aerial photos

Contour interval: 20 meters

0 1 2 KMS

D. B. MacKenzie
Marco 1953

SECTION A–A'



| Sample Locality (See Plate 1) | *Allomorphina* cf. *A. trochoides* | *Alveolina* sp. | *Ammobaculites taylorensis* (?) | *Anomalina* sp. | *Bulimonella lajasensis* n. sp. | *Chofatella decipiens* | *C.* sp. | *Cibicides sukcarinatus* | *Clavulinoides aspera* | *C.* cf. *C. dasjaensis* | *C.* sp. | *Cuneolina* cf. *C. pavonia* | *C.* sp. | *Dentalina calamula* | *D.* sp. | *Dictyomitra* cf. *D. multicostata* | *Discocyclina* sp. | *Discocyclina* sp. | *Discocyclina* sp. | *Dorothia bulletta* | *D. consita* | *D. retusa* | *D.* sp. | *Eponides* sp. | *Gandryina* sp. | *Globigerina*? *cretacea* group | "*Globigerina*" sp. | "*Globigerinella*" sp. | *Globorotalites michelinianus* | *G.* sp. | *Globotruncana* (*Globotruncana*) aff. *G.* (*G.*) *arca* | *G.* (*G.*) *calcarata* | *G.* (*G.*) *calcarata* or *G. ganszeri* | *G.* cf. *G. concavata* | *G. conica* group | *G.* (*G.*) *conica* | *G.* cf. *G. conica* | *G.* aff. *G. coronata* | *G.* (*G.*) *fornicata* group | *G.* (*G.*) *fornicata* | *G.* (*G.*) *fornicata* or *G. conica* | *G.* (*G.*) *fornicata* or *G. conica* | *G.* (*Rugoglobigerina*) cf. *G.* (*R.*) *ganszeri* | *G.* (*R.*) *helvetica* (?) | *G.* (*G.*) *lapparenti* group | *G.* (*G.*) *lapparenti bulloides* | *G.* (*G.*) *lapparenti lapparenti* | *G.* (*G.*) *lapparenti tricarinata* | *G. limestona* | *G. marginata* | *G. marginata* or *G. sardagaensis* | *G.* cf. *G. renzi* | *G.* (*G.*) *rzehaki* group | *G.* (*G.*) *rzehaki* | *G.* (*R.*) *scitula* | *G.* (*G.*) *stuarti* group | *G.* (*G.*) *stuarti* group forma *gr* | *G.* (*G.*) *stuarti* | *G.* (*G.*) *stuarti* or *G. dienstiarti jornis* | *G.* (*G.*) *stuarti* or *G. stigali* | *G. tricarinata* | *G.* sp. | *Gublerina* sp. | *Guembelina* sp. | *G.* aff. *G. globulosa* | *G. plummerae* | *G.* sp. (striate forms) | *Gyroidina beissli* (?) | *Kathina jamaicensis* | *K.* spp. | *Lenticulina* cf. *L. wicha* | *L.* cf. *L. rotulata* | *L.* sp. | *Lepidorbitoides* sp. or *Orbitocyclina* sp. | *Marginulina cretacea* | *M.* sp. | *Marssonella oxycona* | ?*Mandropsina* sp. | *Meyenella meyeni* | ?*Monolepidorbis* sp. | *Neofabellina interpunctata* | *Nodosaria* cf. *N. prolixoides* | *N.* sp. | *Nodosarella* sp. | *Nonionella* sp. | *Omphalocyclus* sp. or *Orbitoides* sp. | *Orbitoides* ss. sp. | Orbitoidal fragments | *Paleonella oralis* | *Planularia* sp. | *Planulina dambisi* | *P.* sp. | *Pleurostomella subnodosa* | ?*Plummerita* sp. | *Polymorphinid* | *Proporocyclina micraulaea* | *Pseudocimolinea* cf. *P. clarata* | *Pseudoglandulina* sp. (?) | *Pseudoglandulina* cf. *P. pygmaea* | *P.* sp. | *Pseudosumbilina costulata* | *P. pseudocostata* | *P. punctulata* | *P. striata* | *Pseudorbitoides israelskyi* | *P.* cf. *P. israelskyi* or *Vaughanina cubensis* | *P.* cf. *P. israelskyi* or *P. rutteni* | *P. rutteni* B-form | *P.* sp. | *Pseudotextularia* sp. | *Pulvinulinella* sp. | *Quinqueloculina* sp. | Radiolaria | Radiolaria or disc-shaped coccoliths | *Rhabdorbitoides kelleri* A | *R. kelleri* B | *Robulus* cf. *R. mamertei* | *R.* sp. | *Rotalispora* cf. *R. appenninica* | *Rugoglobigerina macrocephala* | *R. reicheli hexacamerata* (?) | *R. rugosa* | *R.* sp. | *Spiroplectammina laevis* | Sponge spicules | *Stensioeina* cf. *S. exsculata* | *Subgeneroglina dickersoni* | *S. dickersoni sermunsili* | *S. obesa* | *S.* cf. *S. sermunsili* | *S.* aff. *S. sermunsili* | *S.* sp. | *S.* sp. (evolvect form) | *S.* sp. (large, thin-walled forms) | *Subreinhardtella parida* | *Tergumella* sp. | *Textularia* sp. | *Trochammina* spp. | *Valvulineria* sp. | *Vaughanina cubensis* | *Verneuilina* spp. |
|---|---|---

Probably Late Cretaceous....2189
Late Cretaceous....2244
Conglomerate
Campanian-Maestrichtian (?)....373
Early Cretaceous or younger with reworked Early Cretaceous....445
Cretaceous....505C
Turonian-Campanian....2055
Late Cretaceous, Campanian or younger (pebble)....2120
Campanian (pebbles)....2120
Campanian-Maestrichtian (pebble)....2120
Campanian-Maestrichtian (matrix)....2120
Campanian (matrix)....2120
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MICROFAUNA OF THE MAY AGÜEZ AREA PUERTO RICO

All identifications and age determinations by P. Bronnimann except those marked:
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